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CONTENTS

PAGES

Elongate drift hills of southern Illinois. By John R. Ball..... 951–970

Concentric patterns in the granites of the Llano-Burnet region, Texas. By David Keppel..... 971–1000

Experimental flow of rocks under conditions favoring recrystallization. By David Griggs..... 1001–1022

"Flow" in stressed solids: an interpretation. By Roy W. Goranson.. 1023–1034

Equilibrium between vapor and liquid phases in the system CO_2-H_2O-K_2O-SiO_2. By G. W. Morey and Michael Fleischer..... 1035–1058

Differentiation of the Palisade diabase. By Frederick Walker..... 1059–1106

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

Stratigraphic measurements in parallel folds. By J. B. Mertie, Jr.

Rock series in diabase sills at Duluth, Minnesota. By G. M. Schwartz and A. E. Sandberg

Glaciation in the Wasatch Plateau, Utah. By E. M. Spieker and M. P. Billings

Quaternary history of Louisiana. By Richard Joel Russell

Ep-Archean and Ep-Algonkian erosion surface, Grand Canyon, Arizona. By R. P. Sharp

Rock glaciers in the Colorado Front Range. By Ronald L. Ives

FIGURE 1. LOOKING EAST ALONG ROAD CROSSING COOPS MOUND
Slightly north of the midpoint of the south mound. Geologic section measured along this road beyond the axis of the ridge. Well log obtained from test well at cross roads east of the house.

FIGURE 2. INCLINED BEDDING IN THE GRAVEL OF LARGE PIT
East of the Ware Grove school, north of Butler, Illinois.

COOPS MOUND AND STRUCTURE IN ELONGATE HILL NORTHEAST OF BUTLER, ILLINOIS

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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ELONGATE DRIFT HILLS OF SOUTHERN ILLINOIS

BY JOHN R. BALL

CONTENTS

	Page
Abstract	952
Introduction and acknowledgments	952
The flat drift plain	952
General statement	952
Coops Mound	954
Other elongate drift hills near Coops Mound	957
Robbs Mound	959
Elongate drift hills east of East St. Louis, Illinois	960
General statement	960
Pelican Pouch	961
Elongate hills near New Athens	962
General considerations	962
Pleasant Ridge	962
Belcher Ridge	964
Tamarawa Ridge	965
Elongate hill near Taylorville	966
Salient facts of the elongate drift hills	966
Distribution and general aspects	966
Ground plan and physical features	967
Composition	967
Topographic relations	967
Summary and interpretations	967
General statement	967
Summary of previous investigations	968
Interpretations	969
Résumé and conclusions	970
Works to which reference is made	970

ILLUSTRATIONS

Figure	Page
1. Physiographic divisions of southern Illinois	953
2. Distribution of elongate hills in southern Illinois	954
3. Coops Mound	956
4. Elongate hills north and west of Hillsboro, Illinois	958
5. Robbs Mound	960
6. Pelican Pouch and associated hills	961
7. Pleasant Ridge	962
8. Belcher Ridge	963
9. Tamarawa Ridge	964
10. Elongate hill northeast of Taylorville	965
11. Section of elongate hill northeast of Taylorville	966

Plate	Facing page
1. Coops Mound and structure in elongate hill northeast of Butler, Illinois	951

ABSTRACT

In widely separated areas through southern Illinois are low, elliptical hills of glacial origin. They are in the area of Illinoian glaciation and are somewhat segregated in the western part of the State.

The common occurrence of these hills has instigated recent field studies in southwestern Illinois. The elongate drift hills surmount the uplands and are surrounded by wide expanses of the prairie plains. In some instances they are associated with the knolls of eroded moraines. Their long axes parallel approximately the present drainage lines, and, in one or two instances, they are aligned with pre-glacial drainage trends.

Many of the hills show loess and till in the summit slopes and in road cuts. Others exhibit assorted and stratified drift in deeper road cuts, and certain hills apparently contain gravel cores. In many hills till is abundant, but well logs indicate appreciable thicknesses of gravel. The linear pattern of the hills, their subparallel margins, the position on the uplands, and the presence of gravel or a gravel core, together with other salient features, are considered in the paper as evidence supporting a hypothesis of glacial-fluvial origin for at least some of the hills.

INTRODUCTION AND ACKNOWLEDGMENTS

In widely separated areas through southern Illinois low elliptical hills of glacial material, without bedrock cores, rise above the prairie plains (Fig. 2). These elevations are in the area of Illinoian glacial drift, some of them lying in the flat drift plain and some in the morainic, eroded area (MacClintock, 1935). The purpose of this paper is to assemble the known information regarding them and to offer interpretations.

Dr. M. M. Leighton has read and criticized the manuscript and has made many timely and valuable comments. Originally he offered the suggestion that some of the elongate hills may be of eskerine origin. Dr. G. E. Ekblaw has placed his manuscript map of the moraines of Illinois at the disposal of the writer and has given helpful comments. Drs. G. W. White and Paul MacClintock have read all, and Dr. R. F. Flint has read parts of the manuscript; all have given constructive criticisms. The writer is indebted to Dr. Frank Leverett for information through correspondence, and to Drs. J Harlen Bretz and W. E. Powers for suggestions and criticisms in office conferences. He is indebted to L. E. Workman for the cross section of the hill near Taylorville and for information concerning conditions within the drift ridges.

THE FLAT DRIFT PLAIN

GENERAL STATEMENT

The majority of the hills located and described in this paper are in the flat drift plain. This plain is one of impressive levelness which extends south from the Shelbyville moraine on both sides of the Ridged Drift of the Kaskaskia Basin (Leverett, 1899). It is composed of Illinoian drift, covered by a variable thickness of loess and modified only

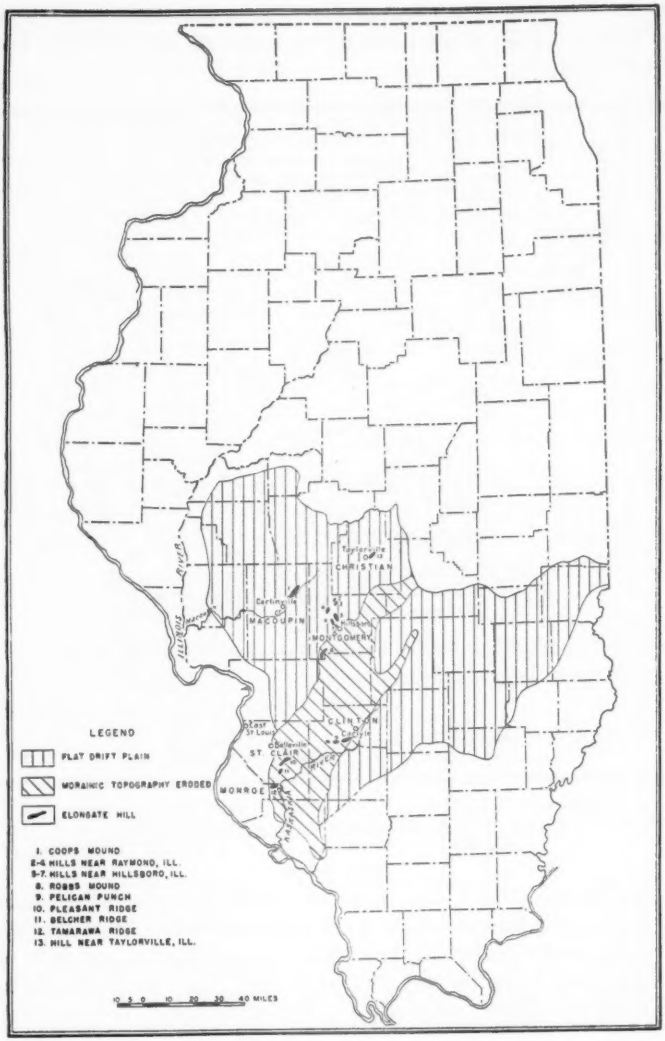


FIGURE 1.—*Physiographic divisions of southern Illinois*
Showing locations of elongate hills. (Modified slightly after MacClintock.)

slightly by erosion. It has been cut into by streams which have developed a dendritic pattern of drainage. The divides are broad and flat, and MacClintock (1929, p. 24) has called this area the "tabular divide country" (Fig. 1).

FIGURE 2.—*Distribution of elongate hills in southern Illinois*
Boundary lines of physiographic provinces after MacClintock.

COOPS MOUND

Coops Mound (Fig. 1, No. 1), in the southeastern part of the Carlinville quadrangle, Illinois, a ridge 2 miles long and with a maximum width of .7 mile, trends N. 29° E., is marked by two elliptical summits (Fig. 3), the highest rising about 60 to 80 feet above the surrounding uplands (Pl. 1, fig. 1). The other summit is farther northeast and is about 40 feet high.

There are several road cuts through the east and south slopes of the mound in sec. 17, T. 10 S., R. 6 W. In a cut on the east side of the mound the following succession of materials is exposed:

		Thickness	
		Feet	Inches
Wisconsin (Peorian):			
	8. Loess, noncalcareous, brown to tan, slumping into gentle slopes	2	
Sangamon:			
	7. Loessial soil, noncalcareous, tan or brown, mottled with gray, nearly vertical; vegetation roots commonly stop on upper surface of this material, hair roots penetrate to its base	1	
	6. Loess-like silt, noncalcareous, deep reddish-brown, with few scattered pebbles of quartz and chert which appear to be secondarily introduced, about	4	6
Illinoian:			
	5. Gravelly concentrate, faintly stratified in its higher levels, somewhat variable in position in section, containing locally manganese or limonite pellets, rounded quartz pebbles, many angular chert fragments, thoroughly leached and oxidized, variable in thickness	0–½	
	4. Till, thoroughly oxidized and leached, gumbo-like, brown, with few large erratics, the largest a gneiss, 12 by 15 inches; occasional pockets of coarse gravel, exposed to bottom of higher cut	4	8

Farther east along the descent of the grade, with its exact relations to the material of the higher cut unknown, and under a covering of slope wash and loess is the following material extending 25 feet along the road:

	3. Gravel, clean, a few angular pebbles up to about 2 inches maximum diameter, larger pieces of chert, the smaller pebbles well rounded	2	6

Along the north-south road at the south end of the mound, in the N½ sec. 20, T. 10 N., R. 6 W., the road cut shows material which is about 30 feet lower than the gravel noted above:

	2. Gumbo gravel, grading upward into slope wash, silt, and loess; medium fine, well-rounded siliceous pebbles	11	

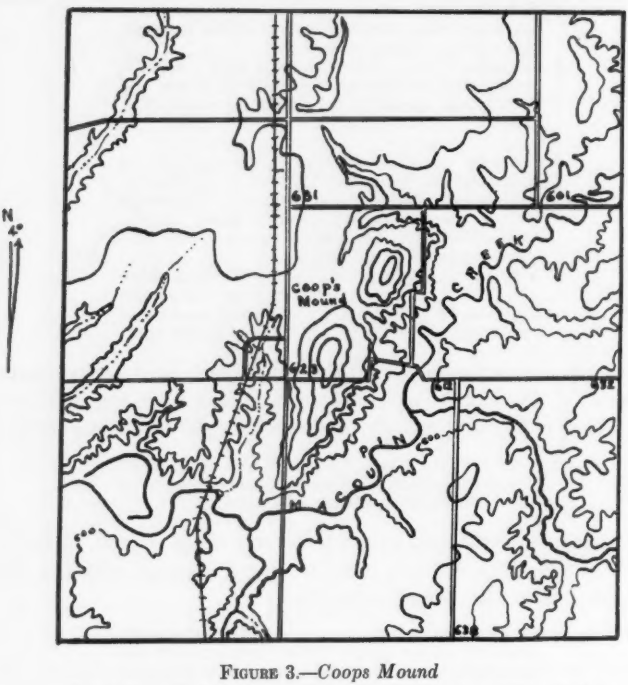
The log of a well in the southwest slope of Coops Mound, NE¼ NE¼ SW¼ sec. 17, T. 10 N., R. 6 W., shows the following:

	1. Gravel, under 18 feet of loess and till	10	
Pennsylvanian:			
	Shale and limestone		

The well is about three-quarters of a mile north of the road cut exposure in the south end of the mound. Its curb elevation is 612 feet above sea level, and its reported gravel [1] is approximately at the elevation of that in the road cut, midway between the house and the flood plain of Macoupin Creek (Fig. 3). About a quarter of a mile west of

[1] Unless otherwise noted, information about well logs has been obtained from the files of the Illinois State Geological Survey, Urbana, Illinois.

the well, west of the railroad and south of the east-west road, a thinner stratum of gravel is exposed in a gully. This gravel is at about the same elevation as that in the well and in the road cut. At the north end of the mound near its axis, a channel-like filling of coarse brown sand, also


N
Coop's Mound
MACOUPIN CREEK
620
600
600
600

major streams of the quadrangle. This parallel relationship of drainage and trend of mound is a common one for eskers.

South and southwest of the mound, not in its projected trend but contiguous to Macoupin Creek, gullies, road cuts, and well logs show that there is considerable outwash in the drift. In well logs from 1½ to 2 miles from the mound, gravel ranging in thickness from 4 to 14 feet is recorded. The gravel is covered by 14 to 18 feet of loess and till. Considerable gravel is exposed in the excavations for Highway 108 across Macoupin Creek 3 miles downstream from Coops Mound. It is in variable thicknesses in the highway cut where it is associated with inclusions of pre-Illinoian till.

Glacial striae within 2½ to 10 miles from the mound record directions of movement S. 40° W. and S. 63° W. The latter reading was taken nearest to the mound which trends N. 29° E., a direction about parallel with the ice movement locally.

OTHER ELONGATE DRIFT HILLS NEAR COOPS MOUND

In the Raymond quadrangle, east of the Carlinville, and in the Mount Olive quadrangle, southeast, are a number of elongate hills (Fig. 1, Nos. 2-7). The hills in the Raymond quadrangle (Fig. 1, Nos. 2-4) are somewhat clustered and are notably smaller than Coops Mound. They have a common alignment—N. 30 E.—which corresponds closely with that of Coops Mound.

The composition of the easternmost hill in this group (Fig. 1, No. 3) is the best known. Under 12 feet of loess the material apparently is fine sand which has been penetrated by a well to a depth of 31 feet. In another well in the hill the sand varies to somewhat coarser texture. In this well the sand is about 19 feet thick. Below the depth of the wells the material in the hill is unknown. Nothing is known about the composition of the other elongate loess-covered hills in this region.

In the Mount Olive quadrangle a group of larger hills, possibly of an end moraine, extends southward from the hills of the Raymond quadrangle to and beyond the city of Hillsboro (Fig. 1, Nos. 5-7; Fig. 4). Their physiography is strikingly similar to that of Coops Mound. Two of the three hills have tandem summits, and the three rise conspicuously above the flat uplands. The hill east of the Ware Grove school (Fig. 4) is slightly more than 1½ miles from the others of the group. The village of Butler is built on the north slope of the largest of the tandem hills. Its trend is continued southeastward in a smaller hill near Hillsboro. The trend of this group of hills intersects that of Coops Mound and that of the hills east of Raymond at a high angle.

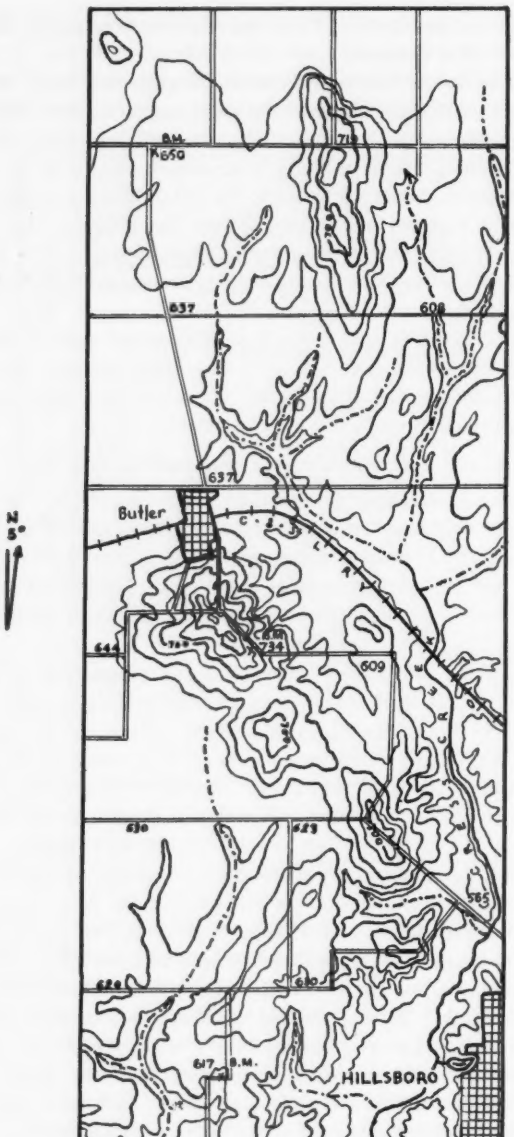


FIGURE 4.—*Elongate Hills north and west of Hillsboro, Illinois*

U. S. Geological Survey, Mount Olive, Illinois sheet (1915). Contour interval, 20 feet; scale, 1 : 62,500.

The somewhat isolated ridge east of the Ware Grove school contains much glacio-fluvial material. Gravel and sand to a depth of more than 30 feet are exposed on the northwest slope of the hill; sand is the more abundant, and some of the beds are inclined. Coarse material is not conspicuous. The gravel extends almost to the top of the hill and the overlying clay apparently is derived from the infiltrated loess (Pl. 1, Fig. 2). Before the gravel pit was excavated, the only gravel visible on this hill was the small amount under the loess in the road cuts. The small exposures are fairly representative of the type displayed elsewhere in the elongate hills.

Fine gravel is exposed on the higher slopes of the hill on which the village of Butler is built. It shows also in the cuts of the road descending this hill easterly along the north line of section 34 (Fig. 4). The hill is from 40 to 60 feet higher than those south of it, and the gravel in it is as high as the summits of the other hills. Till and loess chiefly are exposed in the road cuts of these hills.

Outwash downstream from the hills is exposed in Hillsboro in occurrences similar to those south of Coops Mound. In the northwest outskirts of the city, in a road cut through the bluff of Middle Fork of Shoal Creek, stratified gravel is exposed where a city street joins Highway 16. Gravel and sand are notably cross-bedded. They are at the same level or are incorporated in a gray calcareous till in which the profile of weathering is well developed.

The prominent elongate hills of the Butler-Hillsboro region are associated with much smaller knolls which in some cases also are elongate. The massive elongate hills resemble Coops Mound in ground plan and in their prominence above the surrounding uplands. The highest of the elongate hills is from 120 to 160 feet above the plain. The relief of the highest small elongate knolls is from 40 to 80 feet. This knoll, slightly over half a mile in length, is nearly in line with the larger elongate hills north of it. Most of the minor knolls are of random orientation and are much lower than the smaller elongate knoll.

ROBBS MOUND

Robbs Mound (Fig. 1, No. 8), about 1½ miles northwest of Sorento, trends north-south, is about 1 mile long, and rises to 40 to 60 feet above the uplands. It is the southward continuation of a ridge about as long and slightly higher which has a southwestward trend. The two hills make a pronounced arcuate pattern (Fig. 5).

A small stream is cutting the northwest slope of the northern ridge, and its bed is heavily banked with gravel bars. Near the summit of

Robbs Mound the log of a well shows the following material (Lee, 1926, p. 7):

	Thickness Feet
Illinoian:	
4. Clay	6
3. Gravel	2
2. Sand, coarse, becoming finer below	12
1. Quicksand	2

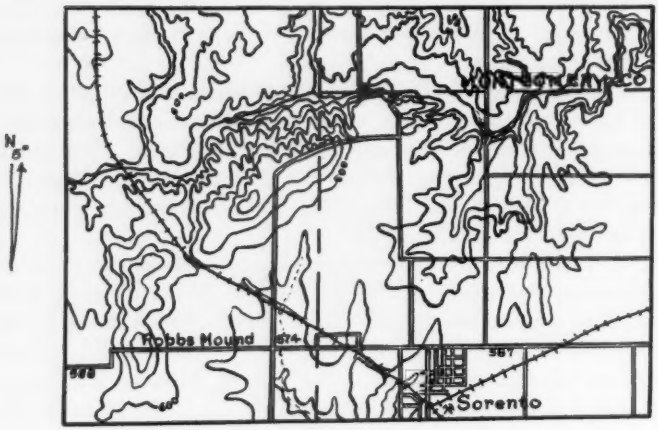


FIGURE 5.—*Robbs Mound*
United States Geological Survey, Mount Olive, Illinois sheet (1915). Contour interval, 20 feet; scale, 1: 62,500.

ELONGATE DRIFT HILLS EAST OF EAST ST. LOUIS, ILLINOIS

GENERAL STATEMENT

Elongate hills are numerous east and southeast of East St. Louis. Those described in this paper lie in the tract called "Topography controlled by morainic drift, eroded" (MacClintock, 1929, p. 23). The hills are distributed through the Belleville, Breese, Carlyle, and New Athens quadrangles, and some receive local names. Most of these hills have been regarded as parts of morainic systems by previous investigators (Leverett, 1899, p. 71-74, and personal communication; Ekblaw, personal communication). The extensive system of ridges which includes the elongate hills of the eroded morainic drift is known as the Ridged Drift of the Kaskaskia Basin (Leverett, 1899, p. 71-72). Certain of these hills resemble Coops Mound in ground plan, and in some of them are noteworthy amounts of cleanly washed gravels.

PELICAN POUCH

Pelican Pouch (Fig. 1, No. 9; Fig. 6) in the Carlyle quadrangle 3½ miles southwest of that village is the largest hill in a group striking east-west. It is about 2.1 miles long and has a maximum width of .7 mile. It trends N. 86° E., a direction nearly parallel with a record of ice

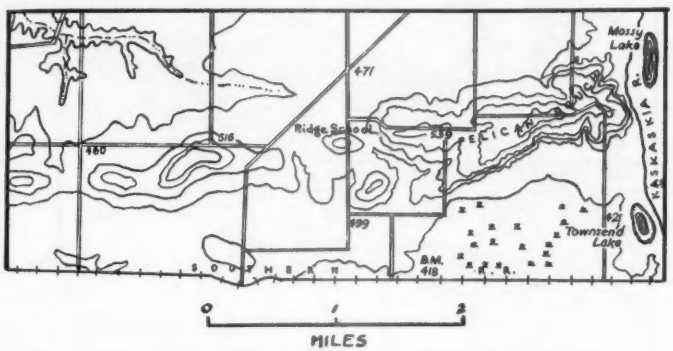


FIGURE 6.—*Pelican Pouch and associated hills*
United States Geological Survey, Carlyle, Illinois sheet (1910). Contour interval, 20 feet.

movement (S. 71° W.). There are three elliptical summits along its crest, the two higher rising from 100 to 140 feet above the uplands. The third summit, farther west, is from 60 to 100 feet high. Pelican Pouch has about the same area as Coops Mound, but its ground plan is slightly different. The east end has been truncated by the Kaskaskia River, and the west end forks.

Gravel occurs high up in Pelican Pouch. In the cut of a secondary road in the SE¼ sec. 2, T. 1 N., R. 3 W., it is exposed about 560 feet above sea level, between 20 and 40 feet below the summit of the ridge. On the slope of the east end of the hill, at a slightly lower elevation, well-rounded pebbles 3 to 5 inches in diameter are surrounded by oxidized sand. Till is exposed in the gullies of the southeast slope at somewhat lower levels, and, still lower, gravel is abundant in the extreme NW¼ sec. 12, T. 1 N., R. 3 W.

Five low hills, continuing the trend of Pelican Pouch, extend about 4 miles farther west. Each of the hills has a slight east-west elongation. The logs of 10 wells, 4 of them approximately on the trend line of the hills, and 3 located on the hills, show sand and gravel ranging from 8 to 114 feet in thickness. With one exception, the greater thicknesses are reported from wells on the trend line. The logs show that the bedrock

is lower along the trend line than it is at the east end of Pelican Pouch. Gravel does not overlie the bedrock in exposures 1 to 3 miles north of the east end of Pelican Pouch. Apparently it is confined to the common trend of Pelican Pouch and its associated hills.

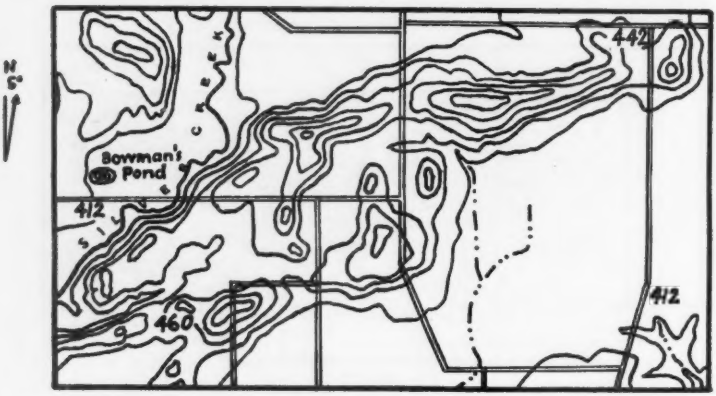


FIGURE 7.—*Pleasant Ridge*

United States Geological Survey, New Athens, Illinois sheet (1912). Contour interval, 20 feet; scale, 1: 62,500.

ELONGATE HILLS NEAR NEW ATHENS

General considerations.—Elongate hills are conspicuous in the New Athens quadrangle. Three of them, in a broken chain across the quadrangle from northeast to southwest, are Pleasant, Belcher, and Tamarawa ridges, respectively (Fig. 1, Nos. 10-12). They are slightly more than 20 miles west and south of Pelican Pouch. Their definite elongation and trend are in strong contrast to other hills in the quadrangle.

The ridges in the New Athens quadrangle are of interest because (1) In definite orientation, ground plan, topography, and composition they resemble closely the elongate drift hills previously discussed, and (2) they are associated with morainic hills (Figs. 7-9).

Pleasant Ridge.—The conspicuous nature of Pleasant Ridge, 3 miles east of Freeburg, is due partly to its isolation (Fig. 7). It is a slightly arcuate ridge, skirted for about 2 miles by Silver Creek which flows S. 56° W. for this distance. It then turns almost directly east for about the same distance. It has the greatest length of the several ridges studied, but its average width is about the same as the others. It has a series of somewhat knoblike summits, although in some of these there is elongation parallel with its trend. Its trend is approximately transverse to the

irregular distribution of scattered morainic knolls which range from northwest to southeast across the quadrangle.

A road cut (SE¼ SW¼ SE¼ sec. 23, T. 1 S., R. 7 W.) in the west slope of Pleasant Ridge shows its composition in some detail. Well-

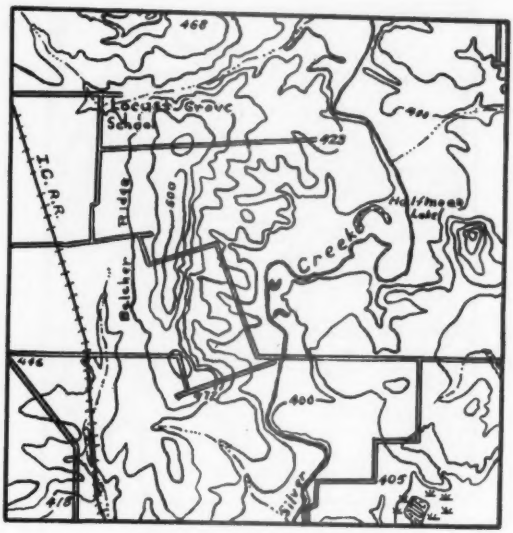


FIGURE 8.—*Belcher Ridge*
United States Geological Survey, New Athens, Illinois sheet (1912).
Contour interval, 20 feet, scale, 1: 62,500.

oxidized gravel, sand, and silt are exposed slightly above the flood plain of Silver Creek. Till and loess overlie the gravel and rise above the uplands. Farther east along the same road about three-quarters of a mile, sand, gravel, and silt are abundant in a cut through one of the low summits.

The log of a well near the east end of Pleasant Ridge in the NW¼ sec. 20, T. 1 S., R. 6 W., gives further data:

		Thickness (Feet)
Illinoian:		
	3. Soil and yellow clay	14
	2. Reddish-brown gravel and sand in layers growing finer with depth	28
	1. Blue clay	83

Road cut exposure and well log both show the composition of Pleasant Ridge along its east-west trend. In some of the hillocks south of the Ridge much of the material is fine sand.

Belcher Ridge.—This is a shorter elongate hill rising slightly above the higher land than that around Pleasant Ridge. Its length along the north-south trend line is about 2½ miles (Fig. 8; Fig. 1, No. 11). There are three knolls along its axis, the middle one elongate, and these are about

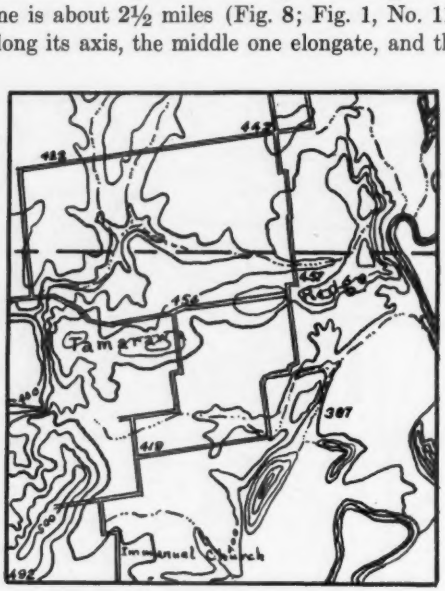


FIGURE 9.—*Tamarawa Ridge*
United States Geological Survey, New Athens, Illinois sheet (1912). Contour interval, 20 feet; scale, 1: 62,500.

40 to 80 feet above the uplands to the west. East of the ridge the upland surface slopes gradually toward the Kaskaskia River and is interrupted locally by morainic hills.

In the cut where a road crosses the south end of Belcher Ridge the upper part of a deposit of medium fine gravel is exposed under 11 feet of loess. This gravel has been eroded to a depth of about 20 feet in a near-by gully. A stratum of fine, water-bearing sand is said to be under the gravel, and a farm well about 100 yards away penetrates gravel and sand for about 40 feet.

Nearly a quarter of a mile further south erosion has exposed the material of the ridge near its axis at levels 40 to 80 feet lower than in the road cut. Here noncalcareous till, about 10 feet thick, overlies fine gravel and fine- to medium-grained sand. Cobbles and boulders up to 14 inches maximum diameter are on the gully floor. It is more probable that the source of the cobbles is in the till than in the gravel.

Tamarawa Ridge.—Tamarawa Ridge (Fig. 9) parallels the eastern extension of Pleasant Ridge. It trends practically at right angles to the general distribution of morainic hills on both sides of the Kaskaskia River. It is about 2¼ miles long and rises from 60 to 100 feet above the up-

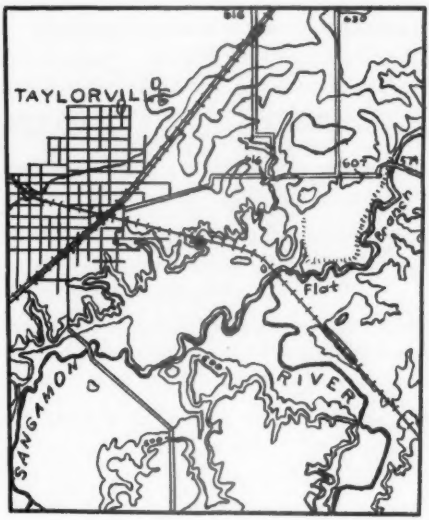


FIGURE 10.—*Elongate hill northeast of Taylorville*

United States Geological Survey, Taylorville, Illinois sheet (1925). Contour interval, 10 feet; scale 1: 62,500.

lands. Its rise above the uplands is less abrupt than in the other ridges of the quadrangle. It has been truncated on both ends by streams. It has five elongate summit knolls, the easternmost partly removed by erosion.

At the base at the east end of the ridge, a core of well-sorted sand and gravel is slumping into the river. The gravel extends for about 100 feet along the stream and rises more than 15 feet above river level.

Slumping obscures the section in the bluff above the gravel, but about a quarter of a mile north the run-off falls into a box canyon possibly 80 feet deep. The measured section in the canyon corresponds somewhat with the structure in the ridge as indicated by well-drilling operations. The canyon walls show the following materials:

		Thickness	
		Feet	Inches
Illinoian:			
	5. Till, calcareous, buff to gray	8–10	
	4. Gravel, rather coarse, well bedded, well assorted	5	6
	3. Till, similar to that above	6	
	2. Sand, medium to fine grain	30	
	1. Talus (till and sand), about	20	

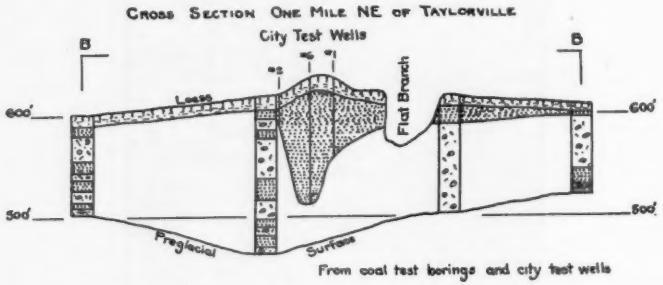


FIGURE 11.—*Section of elongate hill northeast of Taylorville*

By L. E. Workman, Illinois Geological Survey. Published with permission of the Chief, Illinois Geological Survey.

Tamarawa Ridge is said to have a capping of 40 to 45 feet of clay which overlies 140 feet of fine sand. The well did not penetrate ground water, and drilling was abandoned because of caving. The well evidently did not strike the gravel core which probably is nearer the axis of the ridge.

ELONGATE HILL NEAR TAYLORVILLE

One of the least conspicuous of the elongate ridges is located northeast of Taylorville (Fig. 1, No. 13; Fig. 10). Its relief is so slight that a railroad has laid its tracks across it without any conspicuous cuts. It is of irregular outline, slightly arcuate, and extends for some distance beyond the east margin of the Taylorville quadrangle (Leighton, personal communication). Its greatest height near Taylorville is only 20 to 30 feet, but a cross section (Fig. 11) shows that a well near its crest penetrates gravel for nearly 100 feet. The hill is capped by loess only.

SALIENT FACTS OF THE ELONGATE DRIFT HILLS

DISTRIBUTION AND GENERAL ASPECTS

(1) Elongate hills, somewhat resembling Coops Mound, appear south and southwest of the Shelbyville moraine on the Illinoian drift plain, and, as far as determinable from topographic maps, are more numerous in southwestern Illinois.

(2) The general trend of the hills is subparallel to the direction of

movement of the Illinoian ice and is not parallel to the periphery of the ice.

(3) The hills commonly are on the uplands.

(4) Most of the hills are aligned with present drainage. Several of them are close to or in juxtaposition with the valley wall of the stream.

GROUND PLAN AND PHYSICAL FEATURES

(1) The hills are nearly straight, in some instances slightly arcuate.

(2) Some ridges have several almost concordant elongate summit knolls.

(3) The hills are relatively wide in proportion to their length and are much wider than many eskers of Wisconsin age.

(4) Lateral and end slopes are gentle and extended.

(5) Where morainic knolls and elongate hills are associated the latter are larger.

COMPOSITION

(1) The surficial materials commonly exposed in road cuts and high on the slopes are usually loess and till.

(2) In several instances small exposures of fine gravel are in a clay matrix derived from the loess and till. Some of the gravels may be regarded as the upper part of more extensive deposits of stratified drift.

(3) Gravel and sand are exposed in greater quantities in certain hills. The greater amounts of gravel, clearly exposed, are regarded as the gravel core. Evidence of outcrops and well records, taken together, indicate that the cores have considerable linear extent.

TOPOGRAPHIC RELATIONS

The elongate hills are grouped as follows:

(1) Isolated on uplands: Coops Mound, Robbs Mound and the hills east of Raymond.[2] Pelican Pouch probably should be included here, but its axial trend is continued in the small elongate hills westward. Shaw (1923, p. 5) considered the group the nearest approach to the Wisconsin type of moraine in this locality.

(2) Somewhat closely associated with faintly developed morainic hills: the Butler-Hillsboro ridges and the hill northeast of Taylorville.

(3) Definitely associated with morainic hills: Pleasant Ridge, Belcher Ridge and Tamarawa Ridge. Leverett (personal communication) places Pelican Pouch in this group.

SUMMARY AND INTERPRETATIONS

GENERAL STATEMENT

This paper directs attention chiefly to the content of stratified materials in some of the elongate drift hills. The field investigation has covered

[2] Numerous other elongate hills are in this group, but only those mentioned in the paper are named.

so little area that an interpretation commensurate with all the complex glacial problems of the field is not sought at this time. The field conditions have been described in the hope that investigation on the part of others might make possible fuller conclusions concerning the origin and history of the elongate hills. Low hills of similar ground plan in the Wisconsin drift of Ohio have been described as eskers (Hubbard, 1911, p. 72-78). The more massive elongate hills of Illinoian age in Illinois, where not too greatly masked by till and loess, yield striking evidence of glacio-fluvial origin also.

SUMMARY OF PREVIOUS INVESTIGATIONS

The comprehensive studies of Leverett (1899, p. 71-74) led him to postulate the persistence of a lobe in the Kaskaskia Basin after the withdrawal of the ice from adjoining regions. He recognized the anomalous character of the ridges and noted (1899, p. 73) that their trend is nearly parallel with ice movement.

Udden and Shaw (1915, p. 10) stated that none of the ridges show evidence of being strictly interlobate. They favored the suggestion that possibly the ridge-making material was carried near the top of the ice sheet and concluded that the mode of origin of the drift ridges was not established.

Shaw (1923, p. 5) pointed to the peculiarity of the scattered distribution of the hills in the Breese and Carlyle quadrangles and to the fact that they rise abruptly above a flat plain. He was the first to suggest that some of the ridges of the Carlyle and Centralia quadrangles are like "kames or eskers, being composed of sand and gravel and an outside layer of loess".

Lee (1926, p. 9-11), discussing the hills of the Butler-Hillsboro region, states that the conditions which led to the formation of the Kaskaskia drift ridges are not clearly known. He suggested that the broad basin of the Kaskaskia River, probably in existence at the end of Illinoian time, may have had some control of ice movements. He added, however, that the distribution of the Butler-Hillsboro ridges relative to the axis of the Kaskaskia Basin made that control difficult to postulate, and that the question for some time must remain an open one.

MacClintock (1929, p. 23-24) found in southwestern Fayette County, in a region where no elongate hills have been plotted (Fig. 2), that Leverett's hypothesis of a persistent lobe is well supported by the presence of considerable outwash west of the morainic hills. In his study of pre-Illinoian drift in southern Illinois, MacClintock (1929, p. 48-49) noted sand and gumbosand toward the top of one of the Butler-Hillsboro ridges.

MacClintock considered the sand and gumbosand a pre-Illinoian deposit pushed into its present position by later movements of the Illinoian ice.

INTREPRETATIONS

A complete cap of till seemingly exists on some of the hills, and others are covered wholly or in part by loess only. Under these conditions considerations either of tunnel eskers or crevasse fillings arise as tentative hypotheses of origin.

A hypothesis of tunnel eskerine origin does not accord fully with conditions in all the hills because they are discontinuous, relatively short and broad, commonly straight, and distributed over the flat drift plain. Some of them, however, are parallel with present drainage lines, and others are somewhat arcuate. Whether there is any connection or not, the alignment of the elongate hills is the same in some cases as that of the morainic undulations in a near-by marginal moraine. This relationship is evident in the Carlinville and Raymond quadrangles.

The marked elongation and scattered distribution of some of the smaller ridges in the Mount Olive, New Athens, and Breese quadrangles seem suggestive of crevasse filling. This suggestion has force where some of the ridges approach each other almost at 90 degrees. On the basis of other considerations (Leighton, White, personal communications) the hypothesis of crevasse filling obtains further support:

(1) The hills are straight, short, discontinuous, and with subparallel margins.

(2) The extremities of the hills are sharply rounded or pointed in directions both with and against the present drainage (Leighton, personal communication).

(3) The elongate crests, even in the tandem hills, are of uniform elevation, although in some cases one end of the hill may be slightly higher.

(4) The hills rise abruptly from the upland level.

(5) In some of the hills the gravel is apparently covered with loess only.

Certain aspects of some of the hills are suggestive of kamic origin (Leverett, Bretz, personal communications). They are all somewhat massive, and one or two of the hills east of Raymond are quite short. In some of them there is a slight suggestion of a downstream gradient along the crests, and those in which the gravel core is conspicuous are associated with morainic knolls.

If some of the hills are kamic, their elongation might be produced by growth upstream as deposition continued (Bretz, personal communication). Where the hills are on the flat drift plain no other topographic features marking the position of the ice front were developed. The gravels in present downstream relations to some of the elongate hills may indicate

approximately the position of the ice front if they were made by streams issuing from the ice margin. The abrupt deflection of Macoupin Creek at the south end of Coops Mound (Fig. 3) possibly is to be associated with these conditions (Leighton, personal communication).

RÉSUMÉ AND CONCLUSIONS

The elongate hills, both on the flat drift plain and in the morainic eroded area, have the topographic form and the composition of a glaciofluvial deposit.

The elongation and straightness of the hills, their sub-parallel alignment, uniformity of summit elevations, and location on the uplands suggest crevasse fillings. The till cap, if continuous for some of the hills, seems to suggest an eskerine origin for certain of them. In some of the shorter hills, closely associated with the Jacksonville moraine, a kamic origin seems possible.

Further investigation of all the elongate drift hills in southern Illinois is desirable for final determination of all the geological implications. The study summarized in this paper is sufficient only to call attention to the stratified drift content of some of the hills.

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CONCENTRIC PATTERNS IN THE GRANITES OF THE LLANO-BURNET REGION, TEXAS

BY DAVID KEPPEL

CONTENTS

	Page
Abstract	972
Introduction	972
Description of the Buchanan massif	974
General	974
Description of the granites	975
Mineral composition	975
Chemical composition	976
Textural varieties of granite	978
General statement	978
Coarse-grained granite	978
Porphyritic coarse-grained granite	979
Medium-grained granite	979
Fine-grained granite	980
Pegmatite and aplite dikes	980
Contacts with the wall rock	980
Boundaries between textural types	980
Sequence of solidification	981
Primary flow structures	983
Schlieren or flow layers	983
Orientation of xenoliths	983
Parallelism of phenocrysts	984
General mineral parallelism	984
Structural pattern	984
Joint pattern	984
Description of other massifs	985
General statement	985
Kingsland massif	985
Marble Falls massif	987
Enchanted Rock massif	988
Katemcy massif	989
Smoothingiron massif	991
Legion Creek massif	991
Other areas of coarse-grained granite	992
Fine-grained granites of the Llano region	992
Discussion of the mode of intrusion	993
General features of massifs	993
Comparison with other regions	993
Alternate theories of intrusion	994
Proposed theory of intrusion	994

Page

General statement 994
Origin of porphyritic granite 996
Origin of finer-grained central core 996
Summary and conclusions 997
Works to which reference is made 998

ILLUSTRATIONS

Figure Page

1. Location map of massifs studied 973
2. Diagrammatic map and section 974
3. Location of specimens in Tables 1 and 2 976
4. Contact with wall rock 976
5. Dike of porphyritic coarse-grained granite in coarse-grained granite 981
6. Dikes of medium-grained granite cutting coarse-grained granite 982
7. Parallelism of phenocrysts 985
8. Index map of Llano-Burnet uplift 986
9. Quarries at Granite Mountain 988
10. Enchanted Rock massif 989
11. Katemcy massif 990
12. Sections showing suggested earlier stages of typical massif 995

Plate Facing page

1. Textures 978
2. Textures 979
3. Outcrops illustrating structure 982
4. Outcrops illustrating structure 983
5. Buchanan, Kingsland, and Marble Falls massifs 986

ABSTRACT

The larger bodies of pre-Cambrian coarse-grained granites in the Llano-Burnet uplift of central Texas have been studied with special attention to structure and texture. They are steep-walled, roughly circular massifs averaging 12 miles in diameter. Flow layers or schlieren with parallel schist xenoliths are well developed near the borders. Lineation within the flow layers was not detected. The flow layers follow the outlines of the massifs and usually dip outward at angles of 70 degrees or more. Vertical joints normal to the strike of the flow layers are characteristic of the outer portions of the massifs. Pegmatite and aplite dikes near the borders dip toward the centers of the massifs, suggesting marginal fissure fillings.

The granites exhibit a concentric arrangement of three textural varieties. These are, in order of their solidification, an outer coarse-grained granite, an intermediate zone of porphyritic coarse-grained granite, and a core of medium-grained granite. The adjacent textural varieties are mutually gradational indicating simultaneous intrusion. Chemical and petrographic analyses show that the three textural varieties are similar. At a later stage small bodies of fine-grained granite were intruded into the centers of the massifs, around the margins, and in radial dikes.

INTRODUCTION

The Llano-Burnet uplift of central Texas exposes a large area of pre-Cambrian gneiss and schist intruded by granite. The granites range in size from small masses and dikes to batholiths 15 miles or more in diame-

ter, and in texture from fine-grained aplite to very coarse-grained granite. Part of the uplift is described by Sydney Paige (1911; 1912), and the region in general has been discussed more recently by Sellards *et al.* (1932; 1934) and Stenzel (1932). The only structural study of the

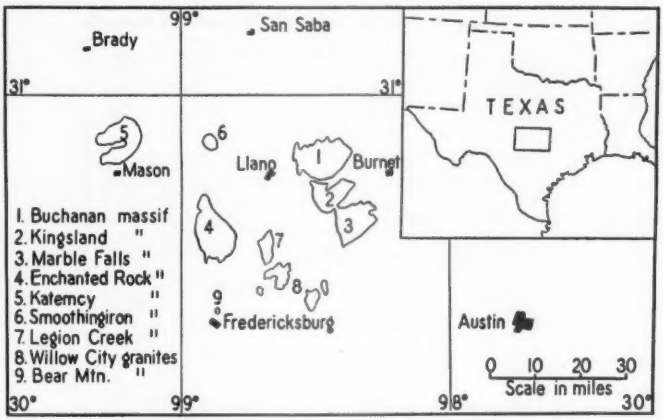


FIGURE 1.—*Location map of massifs studied*

granites is that on the Wolf Mountain phacolith by Stenzel (1936). The present writer has made a study of the internal structural features of the larger massifs of coarser-grained granite (Fig. 1), classed by Stenzel as the Town Mountain granites (Sellards *et al.*, 1934, p. 75).

The typical massif exhibits a concentric arrangement of three textural varieties of granite (Fig. 2) into which small apophyses of a finer-grained granite have been intruded. It is roughly circular in plan view, averaging about 12 miles in diameter, and has steep walls. The different textural phases of granite in each massif are petrographically similar. One area, the Buchanan massif, was studied in detail, and reconnaissance of several others showed repetition of the same general features. Particular attention was paid to the orientation of structural elements within the granite and in the wall rock and to mapping textural varieties.

Helpful discussion with Drs. V. E. Barnes and H. B. Stenzel of the Texas Bureau of Economic Geology has contributed greatly to the work. Dr. J. W. Peoples of Wesleyan University, Dr. S. J. Shand of Columbia University, and Dr. Robert Balk of Mount Holyoke College have read the manuscript and offered valuable criticism. Finally, the writer wishes to acknowledge helpful suggestions of Dr. Paul F. Kerr of Columbia University, under whose guidance the study was made.

DESCRIPTION OF THE BUCHANAN MASSIF

GENERAL

This massif (Pl. 5) is named from Buchanan Lake on the Colorado River which has flooded a large portion of the area since the study was

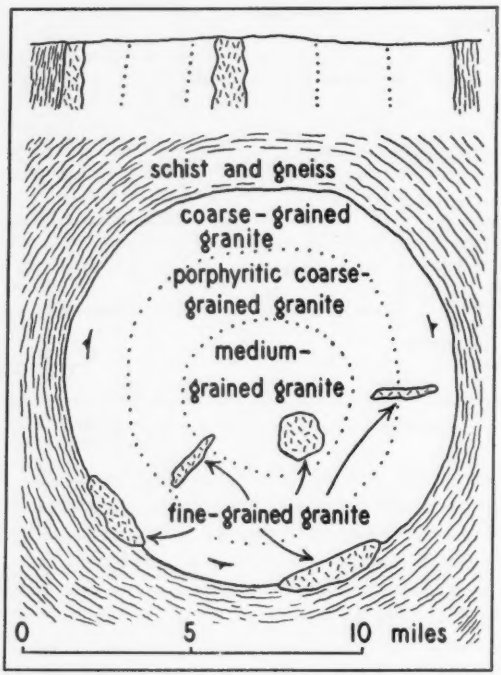


FIGURE 2.—*Diagrammatic map and section*
Showing relationships of textural variations in an idealized massif.

made. It was chosen for detailed study in order to put the flooded portion on the geological record, and because it is a typical example for the entire region. The description of the Buchanan massif could be taken in general to apply to other massifs of the Llano-Burnet region, although each shows individual peculiarities pointed out under the discussion of the other massifs.

The Buchanan massif is roughly circular in outline, about 12 miles in diameter, and bounded by steep walls. The northern border is masked by sedimentary rocks, but enough of the intrusion is exposed to

reveal its principal characteristics. Comparison of the internal structural features of the massif with the borders suggests a rounded outline for the massif to the north, as is indicated by the dotted line underneath the Paleozoic rocks in Plate 5.

TABLE 1.—*Rosiwal analyses*

	Buchanan massif			Enchanted Rock massif			Other massifs		Fine-grained dikes	
	A	B	C	D	E	F	G	H	I	J
Microline	.....	.....	.....	.....	.....	.....	.....	.....	30.4	31.5
Plagioclase	.....	.....	.....	.....	.....	.....	.....	.....	36.8	29.8
Total feldspar	66.7	73.3	73.8	68.2	70.2	76.0	63.7	61.9	67.2	61.3
Quartz	21.3	15.9	19.4	27.3	26.3	20.1	30.9	32.4	30.6	35.7
Biotite	12.0	10.8	6.8	4.3	3.4	3.9	5.4	5.6	1.9	3.0
Per cent of microcline phenocrysts	.....	20	.....	.....	43	7	41	.....	.....	.....
Average grain size (mm.) groundmass	7	7	2	6	6	2	7	6	0.5	0.5
phenocrysts	.....	30	.....	.....	40	25	35	.....	.....	.....

A—Coarse-grained granite, Petrich's quarry, 2 mi. west of Buchanan Dam, Llano Co., Texas (Table 2, analysis A; Pl. 2, fig. 1).

B—Porphyritic coarse-grained granite, road cut (Fig. 6), .65 mi. south of second bridge south of intersection of Highways 29 (old) and 261, Llano Co., Texas (Table 2, analysis B; Pl. 1, fig. 3).

C—Medium-grained granite, from road cut .45 mi. south of Highways 29 (old) and 261, Llano Co., Texas (Table 2, analysis C; Pl. 1, fig. 1).

D—Coarse-grained granite, 2 mi. southwest of Putnam Mountain, Llano Co., Texas (Pl. 2, fig. 2).

E—Porphyritic coarse-grained granite. On Cherryspring Creek, 1.5 mi. southwest of House Mountain, Llano Co., Texas (Pl. 1, fig. 4).

F—Medium-grained granite, 2 mi. east of House Mountain, Llano Co., Texas (Pl. 1, fig. 2).

G—Porphyritic coarse-grained granite, Katemcy massif. Large exfoliated dome, 2.5 mi. west of Fredonia, Mason Co., Texas (Pl. 2, fig. 4).

H—Coarse-grained granite, Marble Falls massif. Granite Mountain quarry 2 mi. northwest of Marble Falls, Burnet Co., Texas (Pl. 2, fig. 3).

I—Fine-grained granite, 4½ mi. north of Buchanan Dam postoffice on the edge of Lake Buchanan, Burnet Co., Texas.

J—Fine-grained granite, 2½ mi. southwest of Roy Inks Dam, Llano Co., Texas.

(*See* Figure 3 for location of specimens.)

DESCRIPTION OF THE GRANITES

Mineral composition.—The rock of the Buchanan massif is a pink, soda-rich biotite granite. The principal minerals, in order of abundance, are microcline and microperthite, quartz, albite-oligoclase, and biotite. Accessories include apatite, ilmenite, sphene, zircon, fluorite, augite, and hornblende.

The microperthite is probably injection or deuteric perthite (Alling, 1932, p. 54). The composition of the plagioclase, based on determinations of the extinction angle in sections showing albite twinning, is between $Ab_{88}An_{12}$ and $Ab_{86}An_{14}$; no great variation from this average

composition was observed. Some of the plagioclase is untwinned and resembles orthoclase, but no true orthoclase was noted.

Rosiwal analyses of specimens taken at different points in the granite areas studied (Fig. 3) are shown in Table 1. Measurements of the

FIGURE 3.—*Location of specimens in Tables 1 and 2*

Dashed lines are boundaries between textural phases. Specimens also shown in Plates 1 and 2.

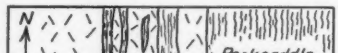


FIGURE 4.—*Contact with wall rock*

East wall of the Buchanan massif, along Morgan Creek, 6 miles north-northeast of Buchanan Dam. Now flooded by the lake.

coarser granites were made on large polished slabs and flat surfaces, and the specific feldspars were not determined. Finer-grained rocks were measured in thin section. In all cases the length of the line measured was over 300 times the grain size. In porphyritic rocks the proportion of phenocrysts was estimated by measuring their intercepts in the same manner as for the mineral proportions. The grain size determinations are averages estimated from hand specimens.

Primarily the analyses show the essential mineralogical similarity of the granites of different textures. Biotite tends to diminish in quantity from the coarser-grained varieties (intruded earliest) to the finer-grained types (intruded last). The proportion of quartz in the granites of the Buchanan massif (A, B, and C in Table 1) is probably too low since the microscopic intergrowths of quartz and feldspar were counted as feldspar in the megascopic measurements.

Alteration of the mafic minerals is most common in the later phases of the intrusion and seems to be primarily deuteric in character. Some fresh specimens of fine-grained granite showed almost complete alteration of biotite to chlorite and of sphene to leucoxene. The feldspars are seldom altered, and in general the minerals are quite fresh, particularly in the coarser-grained granites.

Chemical composition.—The analyses of Table 2 were made to compare the principal textural variations of the Buchanan massif. They are contrasted with a previously published analysis (X) of a quartz

TABLE 2.—*Chemical analyses*

	A—outer zone	B—intermediate zone	C—core	X—later intrusion
SiO_2	72.20	68.36	74.58	75.90
TiO_2	.42	.48	.18	.38
Al_2O_3	13.30	14.82	12.99	12.07
Fe_2O_3	.47	1.26	.67	1.01
FeO	3.28	2.99	1.24	1.45
MnO	.07	.03	.02	n.d.
MgO	.34	.69	.44	.22
CaO	1.10	1.52	.51	.65
Na_2O	3.58	3.88	3.78	3.08
K_2O	4.95	4.98	4.91	5.32
H_2O-	.09	.09	.06	.06
H_2O+	.26	.35	.42	.41
CO_2	none	none	none	none
P_2O_5	.05	.13	.02	.15
S	none	.09	none	n.d.
BaO	.06	.09	.04	n.d.
Cl	none	.02	none	n.d.
SrO	none	none	none	n.d.
	100.17	99.78	99.86	100.70

Recasts of analyses

	A (I,4,2,3)	B (I,4,2,3)	C (I,4,1,3)	X (I,4,1,3)
Quartz	27.07	21.23	31.49	35.72
Orthoclase	29.25	29.47	29.01	31.47
Albite	30.33	32.80	31.96	26.02
Anorthite	5.56	6.67	2.50	2.40
Corundum	.....	.55	.49	.41
Hypersthene	6.05	5.19	2.42	1.74
Ilmenite	.76	.91	.46	.76
Magnetite	.70	1.86	.93	1.39
Pyrite	.....	.18	.....	.....
Apatite	.....	.34	.....	.....

A—Coarse-grained granite, Petrich's Quarry, 2 mi. west of Buchanan Dam, Llano Co., Texas. F. A. Gonyer, analyst. (Table 1, analysis A; Pl. 2, fig. 1.)

B—Porphyritic coarse-grained granite, road cut (Fig. 6) .65 mi. south of second bridge south of intersection of Highways 29 (old) and 261, Llano Co., Texas. F. A. Gonyer, analyst. (Table 1, analysis B; Pl. 1, fig. 3.)

C—Medium-grained granite, road cut .45 mi. south of intersection of Highways 29 (old) and 261, Llano Co., Texas. F. A. Gonyer, analyst. (Table 1, analysis C; Pl. 1, fig. 1.)

X—Opaline quartz porphyry, Llano Co., Texas, quoted in Iddings (1904, p. 228).

(*See* Figure 3 for location of specimens.)

porphyry (Iddings, 1904) which cuts the wall rock and which is probably related to the granites of this massif as a later phase of the same period of intrusion.

The analyses of Table 2 also show the general similarity of the rocks within the Buchanan massif as well as certain progressive changes from

the earliest to the latest rock to crystallize. There is a progressive decrease in hypersthene (represented in the mode by biotite) from the coarse-grained granite (A) on the outside of the massif to the medium-grained granite (C) in the center. Analysis X, the last phase of the granitic intrusion of the Llano-Burnet region, shows still less hypersthene. Quartz increases progressively from A to X with the exception of B, the porphyritic coarse-grained granite which contains a greater proportion of total feldspar. The normative plagioclase computed from the analyses is as follows: for A—$Ab_{85.3}An_{14.7}$, for B—$Ab_{83.9}An_{16.1}$, for C—$Ab_{93.1}An_{6.9}$, and for X—$Ab_{92}An_{8}$, or a change from more calcic on the outside of the massif to more sodic in the center as might be expected. Iddings, in recasting analysis X, took into consideration 0.70 per cent fluorite he observed in the mode. The present writer has recomputed analysis X for better comparison with the other analyses. Fluorite is characteristic of the granites studied in this paper but occurs in minute quantities in the coarse-grained types.

The progressive changes illustrated by the analyses for successive rock types are strikingly similar to the paragenetic relations of the minerals within a single pegmatite dike at the famous Baringer Hill rare earth locality (Hess, 1908), which is now flooded by Lake Buchanan. Here the order of solidification for the minerals, as stated by Landes (1932, p. 388-390), may be summarized as follows:

> The pegmatite was intruded in two phases, magmatic and hydrothermal. The first or magmatic stage included (1) crystallization of microcline, (2) partial replacement of microcline by albite to form perthite, and (3) crystallization of quartz in large masses and replacing microcline to form graphic granite. The second or hydrothermal phase included (1) deposition of fluorite, (2) albitization and the formation of cavities, (3) deposition of second generation quartz, and (4) deposition of the rare earth and accompanying minerals.

TEXTURAL VARIETIES OF GRANITE

General statement.—The Buchanan massif exhibits a wide range in texture. There is an almost continuous gradation from the coarsest to the finest grain size, but certain units were found convenient for mapping. As indicated on Plate 5 the distribution of these textural varieties is not haphazard, but according to a definite concentric pattern with coarse-grained granite on the outside, porphyritic coarse-grained granite in the intermediate zone, and medium-grained granite in the center. Small apophyses of fine-grained granite and dikes of aplite and pegmatite cut into the pattern around the margins and in the center.

Coarse-grained granite.—The grains of this coarse-grained, pink rock (Pl. 2, fig. 1) average about 6 millimeters in diameter, but individual microcline crystals attain 25 millimeters. These larger crystals have

FIGURE 1. MEDIUM-GRAINED GRANITE, BUCHANAN MASSIF

FIGURE 2. MEDIUM-GRAINED GRANITE, ENCHANTED ROCK MASSIF

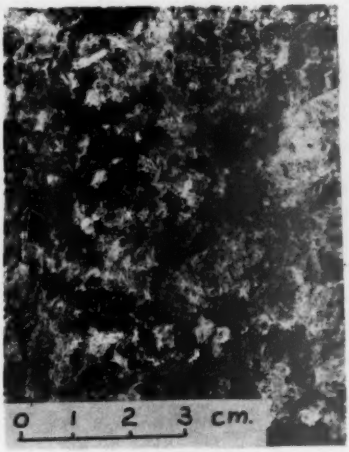


FIGURE 3. PORPHYRITIC COARSE-GRAINED GRANITE, BUCHANAN MASSIF

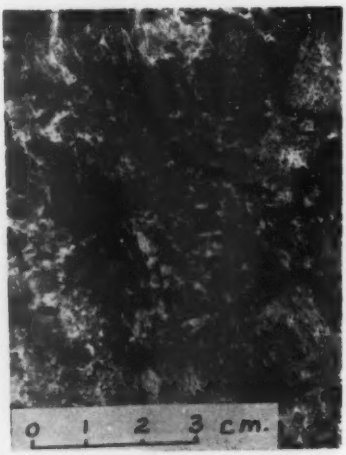


FIGURE 4. PORPHYRITIC COARSE-GRAINED GRANITE, ENCHANTED ROCK MASSIF

TEXTURES

FIGURE 1. COARSE-GRAINED GRANITE, BUCHANAN MASSIF

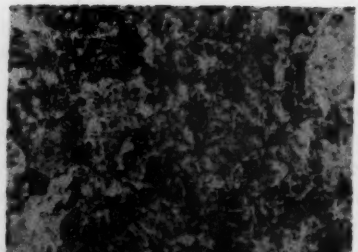


FIGURE 2. COARSE-GRAINED GRANITE, ENCHANTED ROCK MASSIF

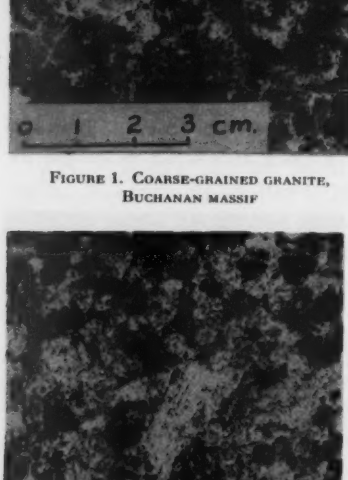


FIGURE 3. COARSE-GRAINED GRANITE, MARBLE FALLS MASSIF

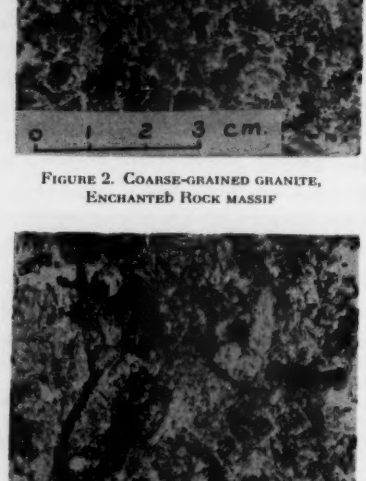


FIGURE 4. PORPHYRITIC COARSE-GRAINED GRANITE, KATEMCY MASSIF

TEXTURES

irregular outlines intergrown with the surrounding minerals and are not phenocrysts.

Coarse-grained granite is quarried in several places and makes a pleasing coarse-grained, pink building stone which is widely used. The Painted Horse Quarry, 2 miles southeast of Lone Grove, and the Texas Quarry, 4 miles west-southwest of the Buchanan Dam, have produced stone in the past. Petrich's Quarry, 2 miles west of the Buchanan Dam, is now in operation.

Porphyritic coarse-grained granite.—This type contains large sharply outlined microcline or microperthite phenocrysts, up to 35 or 40 millimeters long and 15 millimeters thick (Pl. 1, fig. 3). Usually there is every gradation between it and the coarse-grained granite, and sharp contacts cannot be drawn. As pointed out above, the porphyritic coarse-grained granite occupies an intermediate zone, but two small intrusions of this textural type have been mapped in the outer zone of coarse-grained granite near the eastern and western margins of the massif. In both cases the rock contains a greater proportion of biotite than the typical porphyritic coarse-grained granite (Table 1, B). These small bodies of porphyritic coarse-grained granite are not related to the main portion of the intrusion but were probably intruded from outside the massif, where in each case similar rock cuts the wall rock.

The phenocrysts are primarily intratelluric in origin and grew almost to their present size while floating in a liquid. The principal reason for this view is their parallelism which must have been induced by differential movement of the enclosing mass after the phenocrysts had reached nearly full size. Biotite inclusions or a train of random oriented quartz grains paralleling the borders of a phenocryst can best be explained as having been incorporated into the phenocryst during intratelluric growth. Usually, however, the phenocrysts are devoid of inclusions. Where a portion of a biotite crystal from the groundmass is enclosed by the edge of a microcline phenocryst, however, at least some growth of the phenocryst must have been *in situ*.

Medium-grained granite.—The grain size of this variety is distinctly smaller (Pl. 1, fig. 1) than for either of the previous two types, averaging about 2 millimeters. Because of its smaller grain size the medium-grained granite is more resistant to disintegration and has a thinner covering of soil. This difference is useful in distinguishing textures where outcrops are not abundant. Occasional large microcline phenocrysts are found in this variety, but they are never abundant enough for the rock to be called porphyritic. In the Buchanan massif there is no gradation between medium-grained granite and porphyritic coarse-grained granite, but other massifs illustrate every transitional phase between medium-

grained granite with a few 30- to 40-millimeter phenocrysts in a 2-millimeter groundmass to porphyritic coarse-grained granite, with numerous 30- to 40-millimeter phenocrysts in a 6- or 7-millimeter groundmass.

Fine-grained granite.—All the distinctly finer-grained granites with a grain size ranging from 0.5 to 1 millimeter and with no phenocrysts have been classed under this heading. There is less biotite than in the coarser types causing the rock, composed principally of sugary quartz and pink feldspar (Table 1, I), to appear lighter in color. The fine-grained granite exposures stand out prominently in contrast to the more readily weathered coarse-grained rocks.

Pegmatite and aplite dikes.—Pegmatite and aplite dikes, or pegmatite and aplite mixed in the same dike, are seen in most outcrops of the coarser-grained granites. Often the aplitic and pegmatitic phases are segregated in discontinuous layers or lenses parallel to the wall of the dike. The dikes range from less than an inch to more than 100 feet in width. In the fine-grained granite pegmatites are less abundant; they are lenticular bodies, about 5 feet long and a foot wide, without sharp boundaries, and parallel to the flow structure of the surrounding rock. They seem to be concentrations of volatile material which have operated to change the texture locally.

The occurrence of rare earth minerals at the Baringer Hill pegmatite (Hess, 1908; Landes, 1932; Barrell, 1917; Holmes, 1931) has been mentioned. Discoveries of unusual minerals have been largely confined to Baringer Hill, and only traces are found in other pegmatites.

CONTACTS WITH THE WALL ROCK

Wherever exposed the contact is sharp and distinct. Figure 4 shows the nature of the contact on the east wall of the massif. Similar exposures may be seen on the south and west walls. The northern end of the massif is covered by Paleozoic sediments, and the contact is not exposed. Transitional phases between granite and the Packsaddle schist are lacking, and there is no evidence for granitization of the schist. The contact with the Valley Spring gneiss is not exposed in the Buchanan massif, but it is a sharp break in other massifs.

BOUNDARIES BETWEEN TEXTURAL TYPES

As mentioned, the outer zone of coarse-grained granite is transitional to the intermediate zone of porphyritic coarse-grained granite. One passes from typical coarse-grained granite through a zone where large microcline crystals become progressively more euhedral in outline. At this point the rock may be classed as porphyritic coarse-grained granite.

There is no sharp break; the boundary shown on the map is the halfway point in the transition. The contacts of the two small intrusions of porphyritic coarse-grained granite with coarse-grained granite on the east and west margins of the massif, respectively, have been shown as

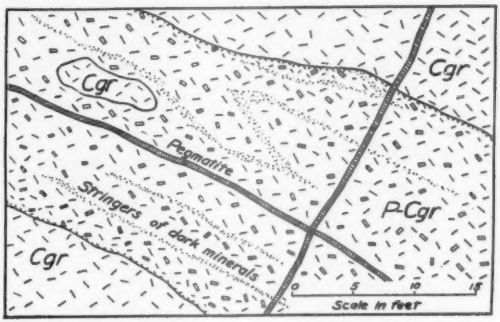


FIGURE 5.—*Dike of porphyritic coarse-grained granite (P-Cgr) in coarse-grained granite (Cgr)*

Taken from a field sketch. Now covered by Lake Buchanan, 2 miles southeast of Tow.

a sharp break, because in each case adjoining outcrops are quite different, although no single outcrop could be found showing the actual break.

The central core of medium-grained granite is separated by a sharp break from the surrounding porphyritic coarse-grained granite. There is an abrupt change in textural type on either side of the line, with no intermediate phases. At two points along the contact, 1½ miles west-southwest and 3 miles northwest of Bluffton, the contact is marked by an unusually thick accumulation of pegmatite dikes. Both these points mark projections of the medium-grained granite mass into the surrounding rock and may have been favorable locations for the accumulation of pegmatite material along the contact.

The contact of the fine-grained granite bodies with the surrounding rock is always sharp and may be seen in a number of places. Elsewhere the contrast in weathering characteristics is so marked that a definite boundary can be drawn.

SEQUENCE OF SOLIDIFICATION

The order of solidification of the granites and related rocks of the Buchanan massif is as follows: (1) coarse-grained granite, (2) porphyritic coarse-grained granite, (3) medium-grained granite, and (4) fine-grained granite. Aplite and pegmatite dikes are later than the rocks

in which they occur but are probably of different ages in different places. For example, a pegmatite dike in the coarse-grained granite might be older than the medium-grained granite, it might be of the same age having been derived from the latter, or it might be younger than either.

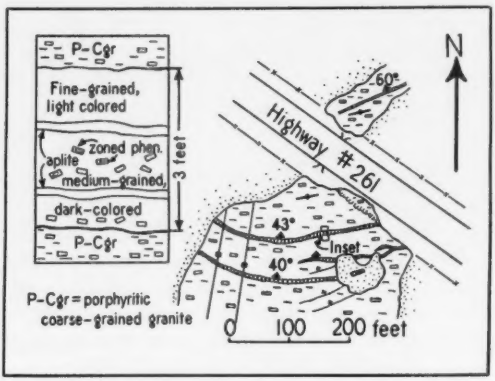


FIGURE 6.—*Dikes of medium-grained granite cutting coarse-grained granite*

Inset shows detail of dike at point marked on base map. Along highway 4 miles northwest of Buchanan dam.

Although the order given is what one would normally expect (solidification of the outer portions of the intrusion first and the center later), positive evidence of the sequence must be given, since the reverse sequence is possible.

In the northern part of the massif, 2 miles southeast of Tow, an outcrop of coarse-grained granite (Fig. 5), now beneath Lake Buchanan, is cut by a 25-foot dike of porphyritic coarse-grained granite with biotite schlieren paralleling the contact, and enclosing a block of the wall rock (Pl. 3, fig. 3). Although several microcline phenocrysts occur close to the contact, none are sheared off at the contact, and the groundmass is noticeably finer close to the margin, eliminating the possibility that the dike might be a xenolith. Relations at this outcrop and elsewhere indicate that the porphyritic coarse-grained granite is later than the coarse-grained granite.

Medium-grained granite cutting porphyritic coarse-grained granite is exposed on the new highway along the west shore of Lake Buchanan about 4 miles northwest of the dam (Fig. 6). Dikes of medium-grained granite, mixed with aplitic bands and containing phenocrysts parallel

FIGURE 1. SCHIST XENOLITH

FIGURE 2. MEDIUM-GRAINED GRANITE DIKE
Cutting porphyritic coarse-grained granite.

FIGURE 3. BLOCK OF COARSE-GRAINED GRANITE (UNDER HAMMER)
In porphyritic coarse-grained granite.

FIGURE 4. PARALLELISM OF PHENOCRYSTS
In porphyritic coarse-grained granite.

OUTCROPS ILLUSTRATING STRUCTURE

FIGURE 1. CONTACT OF GRANITE AND SCHIST

FIGURE 2. SCHLIEREN, GRANITE MOUNTAIN QUARRY

FIGURE 3. TYPICAL OUTCROP OF COARSE-GRAINED GRANITE

FIGURE 4. EXFOLIATION DOME
In coarse-grained granite. Kennison Rock, Burnet County.

OUTCROPS ILLUSTRATING STRUCTURE

to the border, dip at about 40 degrees toward the medium-grained granite of the core, 1.5 miles to the north. A small intrusion of medium-grained granite in porphyritic coarse-grained granite is exposed about a mile west of the core along the highway. These facts, together with the sharp boundaries of the medium-grained granite, indicate that it solidified after the coarser varieties and probably was intruded at a slightly later date in the same period of intrusion.

Fine-grained granite cuts all other textural varieties, but it is not known whether the fine-grained granite was intruded during the same period of granite injection or whether it belongs to a later period. It is overlain unconformably by Cambrian formations and so must be pre-Cambrian. Its position around the edge of the larger intrusion and its petrographic similarity to the earlier granite suggest that it was intruded after medium-grained granite but during the same period of intrusion. Near the northeast margin of the Buchanan massif, just south of the line of section *A-A'* in Plate 5, fine-grained granite of two ages may be observed: An intrusion with grain size averaging 1 millimeter forms masses following the contact of the coarse-grained granite with the wall rock and is cut by small masses with grains averaging 0.5 millimeters and intruding the massif radially.

PRIMARY FLOW STRUCTURES

Schlieren or flow layers.—At good exposures and in quarries streaks containing more abundant dark minerals are seen to be the outcrops of schlieren or flow layers (Pl. 4, fig. 2). They are principally developed near the edges of the massif and are infrequent in the core. Sometimes discontinuous sills of aplite, not more than a foot wide, are parallel to flow layers. It is suggested that these are early segregations derived from the consolidating magma and are essentially contemporaneous with the formation of primary flow layers.

Numerous dark streaks on the surface of flat outcrops (Pl. 4, fig. 3) classed under general mineral parallelism, to be described, are probably flow layers also, but their exact nature cannot be proved owing to the lack of a third dimension.

Orientation of xenoliths.—Xenoliths (Pl. 3, fig. 1) are common in the border zones. They are often found near contacts with the Packsaddle schist and more rarely near contacts with the Valley Spring gneiss. Where abundant, they usually have their long axes parallel, conforming to other structures, such as flow layers or phenocrysts. Orientation of xenoliths is one of the most reliable and sometimes the only method of detecting flow structure.

OUTCROPS ILLUSTRATING STRUCTURE

Parallelism of phenocrysts.—In porphyritic coarse-grained granite, and sometimes in medium-grained granite where phenocrysts are abundant, the larger microcline phenocrysts are commonly oriented in parallel planes (Pl. 3, fig. 4). Schlieren frequently parallel this plane. No linear parallelism of the phenocrysts could be detected, even where surfaces at right angles are available for study. This situation is illustrated in Figure 7—parallelism in two directions with random orientation in the third.

General mineral parallelism.—Under this head are included all structural features of the granites which can be seen on one surface only and which cannot be identified definitely with any of the above primary structural features. In most cases the large flat outcrops of granite (Pl. 4, figs. 3 and 4) exhibit a mineral parallelism which can be detected by comparing several observations. Two types of such parallelism have been recognized: one for phenocrysts, and another for rocks which are not porphyritic. In the latter case orientation of numerous small trains of biotite, taken in conjunction with a spindle-shaped elongation of the quartz and feldspar, gives a sort of "grain" to the rock.

Structural pattern.—The Buchanan massif is bounded by steep walls exposed on the east, west, and south sides. Flow layers also dip steeply and parallel the walls. If the contact is projected underneath the overlapping Paleozoic sedimentary rocks to the north conformable with the flow layers, the massif is roughly circular in outline as indicated on Plate 5. The textural varieties of granite form a concentric pattern which follows the outlines of the structural pattern. Thus the massif consists of concentric shells of textural units (Fig. 7), all portions of which dip steeply. It might be classed as a schlieren dome (Balk, 1937, p. 56), which has been so deeply eroded that the arching of flow layers over the central portion has been completely removed.

Joint pattern.—Many joints were probably formed during the consolidation of the rock because they are related to the structural pattern. In the outer portions of the massifs, vertical or steeply dipping joints strike approximately at right angles to the flow layers in a fan-shaped pattern. These are sometimes accompanied by a minor set of steep or vertical joints at right angles, or nearly parallel to the flow layers, but only the major joints are mapped.

The arrangement of elongated bodies of fine-grained granite near the eastern margin of the Buchanan massif in a fan-shaped pattern suggests that their intrusion was controlled by the primary joint pattern already present in the outer margins of the massif. Joint systems in dikes and small intrusive bodies of fine-grained granite are usually

distinct from the systems in the surrounding coarser granites, showing that they are due to stresses formed during the consolidation of the dike.

Flat-lying or gently dipping joints of primary origin are not common, though their detection is made more difficult by the flat nature of the

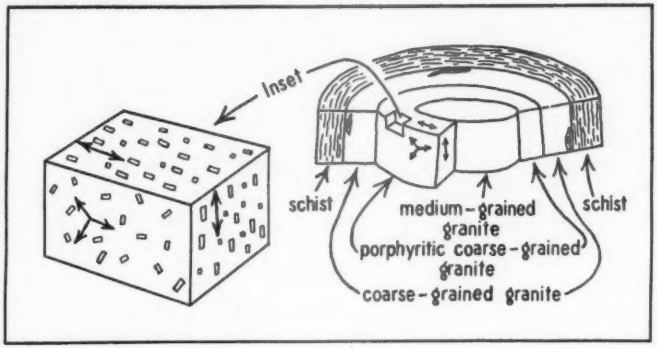


FIGURE 7.—*Parallelism of phenocrysts*

In porphyritic coarse-grained granite. Diagrammatic sketch showing parallel orientation in two directions, with random orientation in the third. No linear parallelism of long axes is present.

outcrops. Gently dipping dikes (Fig. 6) probably were guided by joints of this nature.

Other joints were formed long after the consolidation of the granite and can be ascribed to special causes. Near known Paleozoic faults closely spaced joints in a rather brecciated granite frequently parallel the fault. Where such closely spaced joints appear far from Paleozoic rocks, they may be associated with Paleozoic faulting, the evidence for which has been eroded away. The significance of still other joints is not known.

DESCRIPTION OF OTHER MASSIFS

GENERAL STATEMENT

The areas here described include the major occurrences of the coarser-grained granites in the Llano-Burnet region (Fig. 8). The petrographic characteristics of the granites, textural varieties, and types of flow structures are so similar to those of the Buchanan massif that no description is given for them. Only those features which supplement the phenomena described for the Buchanan massif are mentioned.

KINGSLAND MASSIF

The Kingsland massif (Pl. 5) does not show the complete set of concentric rings, although portions of bands of the three textural varieties

may be traced. Furthermore, it is not completely surrounded by schist but merges with the Marble Falls massif on the southeast, and the schist septum to the north is quite thin, suggesting that it may be connected at depth with the Buchanan massif.

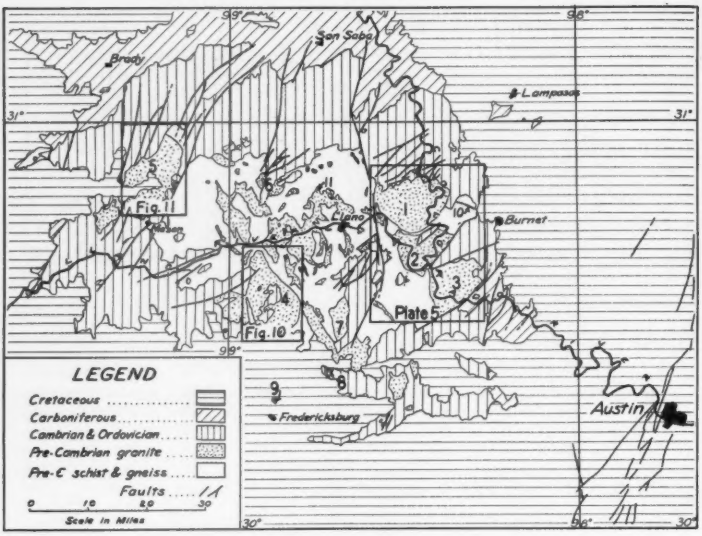
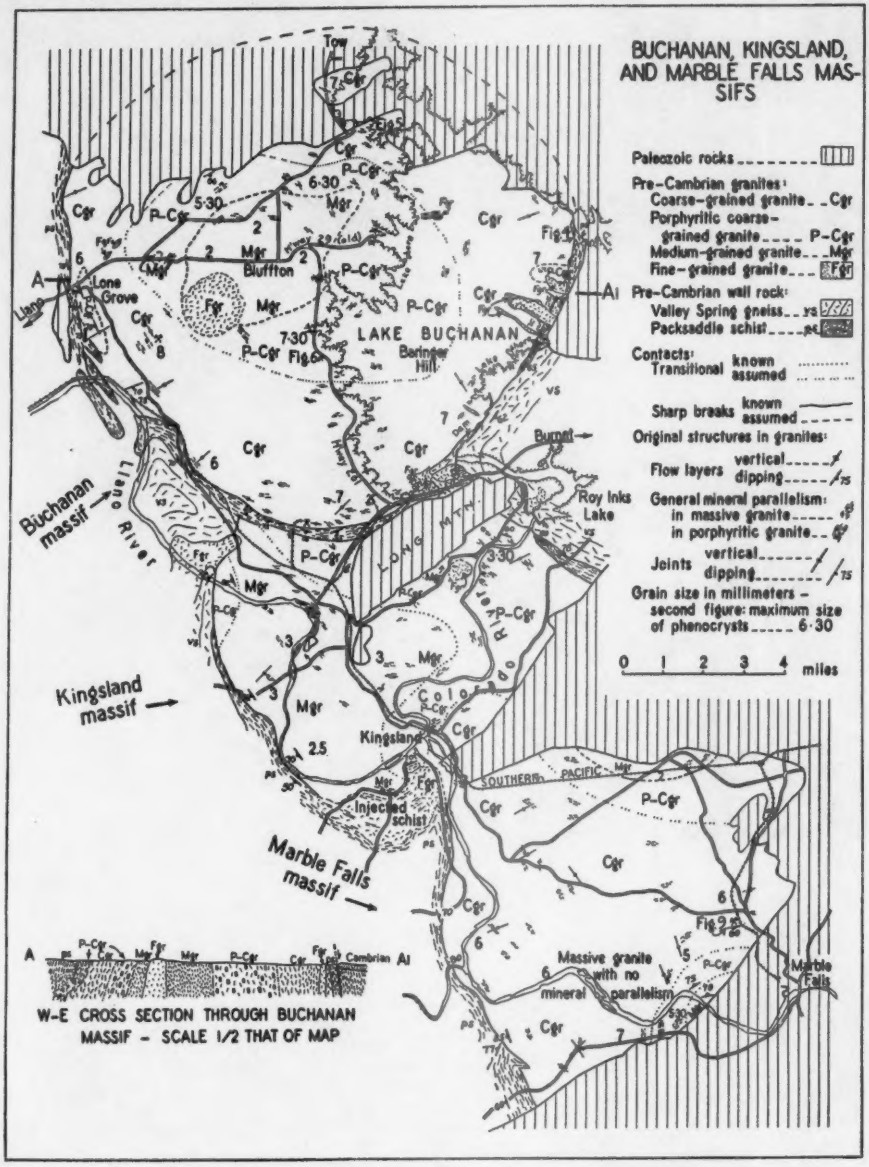


FIGURE 8.—*Index map of Llano-Burnet uplift*

Showing location of Plate 5 and Figures 10 and 11. Generalized from geologic map of Texas, scale 1/500,000 (U. S. G. S., 1937). 1. Buchanan massif, 2. Kingsland massif, 3. Marble Falls massif, 4. Enchanted Rock massif, 5. Katemcy massif, City granites, 9. Bear Mountain granite, 10. Midway sill, 11. Wolf Mountain phacolith.

There are two cores of medium-grained granite: a smaller northern one, and a larger southern one. The latter becomes broader to the west and adjoins the marginal schist for a considerable distance along the border without the usual phases of porphyritic coarse-grained granite and coarse-grained granite. In each case there is every transitional phase between medium-grained granite and porphyritic coarse-grained granite. Fine-grained granite intrudes along the northwestern border, in a mass in the north-central portion (Table 1, J), and along the southern border.

Blocks of an earlier porphyritic granite about 3 feet on a side containing schist xenoliths occur in porphyritic coarse-grained granite along a stream bed 1¼ miles S. 80° W. of Roy Inks Dam. The earlier porphyritic granite has microcline and microperthite phenocrysts about



BUCHANAN, KINGSLAND, AND MARBLE FALLS MASSIFS

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10 mm. long and quartz, microcline, albite-oligoclase, and biotite averaging 3-4 mm. in the groundmass. The minerals of the groundmass have been injected along crystal boundaries and largely replaced by quartz granules 0.1 mm. in diameter. The surrounding porphyritic coarse-grained granite is typical for the region. The locality is half a mile from the margin of the massif, and there are numerous schist xenoliths near by. The blocks appear to represent the remnants of an earlier granite intrusion stoped off and partially replaced by the main body of the massif.

MARBLE FALLS MASSIF

An unknown, but presumably large part of this massif (Pl. 5) is not exposed, since the edge of the granite on all but the western side is covered by Paleozoic rocks. There are two areas composed of medium-grained granite, one in the northern part and another in the southern part. The curvature of the exposed portions of their contacts suggests that they are the edges of cores, and it is therefore believed that the granite area probably extends north and south underneath the Paleozoic rocks for some distance. A large area of coarse-grained granite between these cores, and slightly to the west, shows poor flow structure and is quite massive. Extensive outcrops in the vicinity of Granite Falls on the Colorado River are completely devoid of primary structures, and even close to the contact with the schist to the west flow layers are poorly developed. Scattered outcrops within the area mapped as coarse-grained granite are quite porphyritic but, since they could not be separated into definite bands, they have been included with the dominant rock type.

The quarries at Granite Mountain (Fig. 9), 2 miles northwest of Marble Falls, expose a remarkable series of parallel pegmatite and aplite dikes, striking east-west and dipping 30°-40° N. They range in width from 6 inches up to 10 feet or more, and the largest one contains molybdenite and some of the rare earth minerals.

The quarries have been in operation for many years and have produced large quantities of stone (Pl. 2, fig. 3), the greater proportion of which is in the jetties and sea walls of the Texas coast. Several exposures on the quarry walls exhibit schlieren (Pl. 4, fig. 2), which are consistent in general trend for the mountain as a whole. Joint systems striking over the top of the mountain are the controlling factors for the breaking of the rock in many places.

Along the southern part of the western edge of the Marble Falls massif, pegmatite dikes dip at angles of 30 to 40 degrees away from the contact. These may be marginal fissures (Balk, 1937, p. 101). Since the actual contact was not exposed, their continuation into the wall rock could not be demonstrated.

ENCHANTED ROCK MASSIF

In the southwestern portion of Llano County a large massif is named after Enchanted Rock, a perfect exfoliation dome 160 acres in area and one of a series of similar domes near the eastern edge of the massif

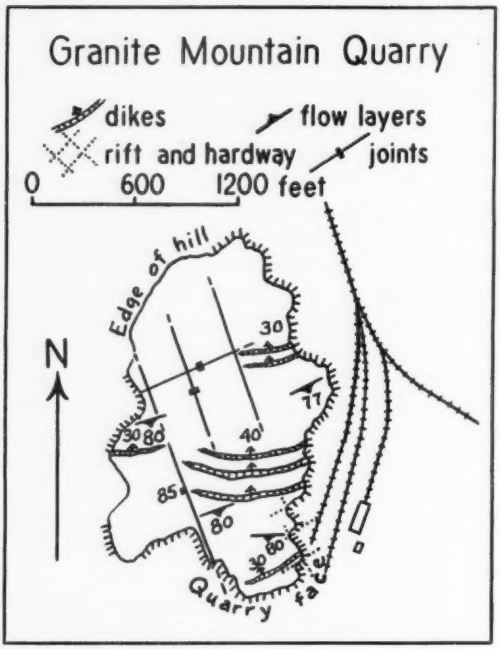

FIGURE 9.—*Quarries at Granite Mountain*

2 miles northwest of Marble Falls, Burnet Co., Texas.

(Fig. 10). The contact with the wall rock is exposed on all but the southern side, and a relatively small proportion of the surface is covered by later sedimentary formations. The massif illustrates well the concentric zonal distribution of textural phases of granite as described in the Buchanan massif. In this case the marginal rim of coarse-grained granite (Pl. 2, fig. 2) is narrower and in places absent. There is a well-developed central core of medium-grained granite (Pl. 1, fig. 2) which everywhere contains large phenocrysts and which is transitional to the intermediate ring of porphyritic coarse-grained granite (Pl. 1, fig. 4). Two areas of fine-grained granite are recognized. As in the case of the

Buchanan massif, the contacts with the wall rock are steeply dipping. Finally, vertical joints normal to the flow layers around the edges spread out radially from the center of the massif.

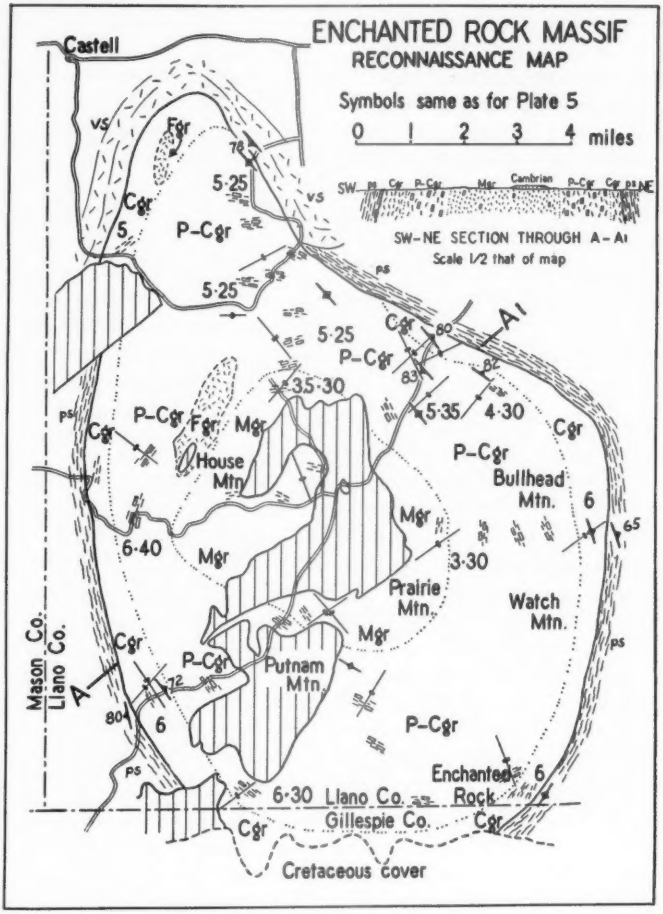

FIGURE 10.—*Enchanted Rock massif*

KATEMCY MASSIF

Named from exposures near the town of Katemcy in northern Mason County, this massif (Fig. 11) occurs in a region that has never been

mapped accurately geologically, and the boundaries shown are generalized from a reconnaissance survey. The general picture is that of a granite massif, roughly circular in outline, the western half of which is

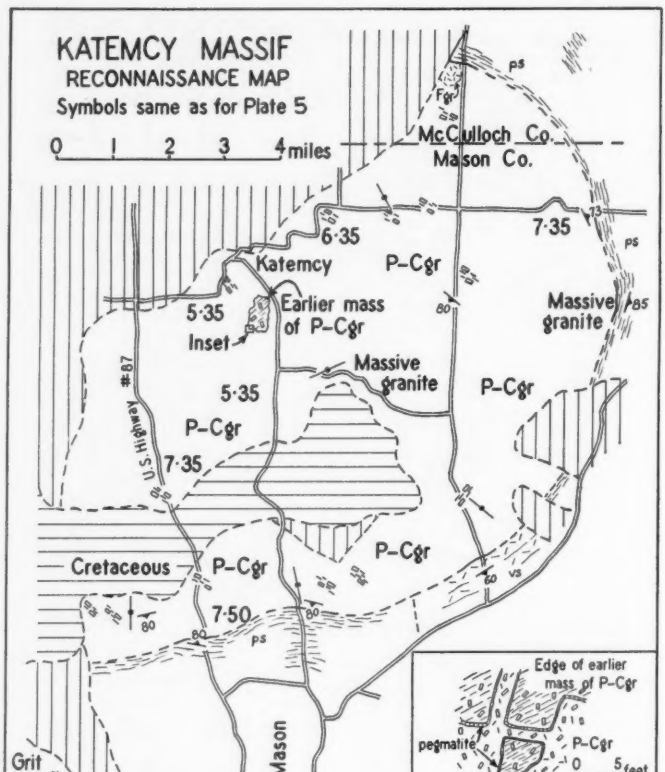

FIGURE 11.—*Katemcy massif*

covered by sedimentary rocks. The whole area exposed has been mapped as porphyritic coarse-grained granite (Pl. 2, fig. 4). If an outer rim of coarse-grained granite is present, it must be very narrow, as none was seen in any of the traverses made. No core of medium-grained granite was discovered, although one may exist to the west beneath Cretaceous rocks. North-dipping dikes of medium-grained granite (Pl. 3,

fig. 2), similar to those shown in Figure 6, cut the porphyritic coarse-grained granite south of the most easterly extension of Cretaceous rocks shown on the map. Flow structures follow the outlines of the massif, and vertical joints are normal to flow layers. The central and eastern parts of the exposed area are composed of massive rocks with no flow structures.

Porphyritic coarse-grained granite outcrops south and east of Grit, southwest of the Katemcy massif. This may or may not be connected with the larger mass, as it is separated by sedimentary cover.

Granite earlier than the principal rock composing the massif crops out a mile south of the town of Katemcy. A considerable area of porphyritic granite with a finer groundmass than the porphyritic coarse-grained granite which intrudes it resembles the blocks of earlier granite described in the Kingsland massif. (*See* inset of Figure 11 showing nature of the contact.) A similar rock was noticed about 6 miles south of the contact with the wall rock along the Mason-Brady highway. This earlier porphyritic granite lacks parallelism of the phenocrysts, suggesting that they may have formed *in situ*.

SMOOTHINGIRON MASSIF

Named from Smoothingiron Mountain between Valley Spring and Field Creek in northwestern Llano County, this is the smallest massif studied—about 4.5 miles in diameter. The granite is texturally homogeneous—a coarse-grained granite, not noticeably porphyritic, but finer-grained than the usual coarse-grained granite of this region (averaging 3 or 4 millimeters as opposed to 6 or 7 millimeters). No porphyritic coarse-grained granite or medium-grained granite phases were seen. Flow structure is indistinct, and the rock is quite massive. Although about half the area of the massif is covered by Cambrian sedimentary formations, scattered exposures suggest that the granite in the covered portion is texturally the same as the rest. Since this massif is less than half the diameter of the others studied, it may have been too small for the development of the usual concentric rings, or it may be the upper portion of an intrusion which at greater depths would show the textural variations of the larger massifs.

LEGION CREEK MASSIF

In south-central Llano County, and extending into Gillespie County, an area of about 20 square miles of granite is exposed south of Sandy Creek and west of Cedar Mountain. The eastern half of the massif, named from Legion Creek, a tributary of Sandy Creek, is covered by the Paleozoic rocks of Cedar Mountain. Reconnaissance of the area shows that most of the rock is coarse-grained or porphyritic coarse-

grained granite but that finer-grained material is exposed just west of Cedar Mountain about a mile north of the Gillespie-Llano County line. This is about the center of the massif (if the covered eastern portion were restored). Thus it appears probable that this massif would also show a concentric pattern, with finer-grained material in the center.

OTHER AREAS OF COARSE-GRAINED GRANITE

Coarse-grained granites occur at several other localities in the Llano-Burnet region, and doubtless great expanses of them are covered by Paleozoic and later rocks. Some of these are of the batholithic type already described, whereas others have been shown to be concordant intrusions.

In northeastern Gillespie and northwestern Blanco counties coarse-grained granite areas surrounded by sedimentary rocks are apparently portions of a much larger batholith. Porphyritic coarse-grained granite is exposed just west and several miles north of Willow City in Gillespie County, and some of the coarsest porphyritic granite seen anywhere in the region is exposed north and west of the town of Sandy in Blanco County. Coarse-grained granite with some medium-grained granite crops out northeast and east of Art in eastern Mason County over a considerable area.

Concordant or floored masses of the coarse granites have been described. Northwest of Llano Stenzel (1936) and McAdams (1936) have described a phacolith intruded along the contact of the Valley Spring gneiss and the Packsaddle schist. The rock is a porphyritic granite, slightly finer-grained and darker than the usual porphyritic coarse-grained granite from the larger massifs and quite similar to the two small intrusions of porphyritic granite on the eastern and western margins of the Buchanan massif. The Midway sill (Sellards and Baker, 1934, p. 75, 77) is a steeply dipping sill of porphyritic coarse-grained granite 3½ miles long and less than a quarter of a mile wide northeast of the Buchanan massif.

Four miles north of Fredericksburg at Bear Mountain a coarse-grained granite has been quarried for many years. It outcrops in a small knob completely surrounded by Cretaceous rocks. The rock is not at all porphyritic, contains a very small proportion of dark minerals giving it a pinker aspect than the other pink granites, and the average grain size is about 4 millimeters. Flow structure is notably absent. The quarry shows considerable jointing, although large blocks can be taken out.

FINE-GRAINED GRANITES OF THE LLANO REGION

The schists and gneisses of the Llano-Burnet region are extensively intruded by small masses and dikes of fine-grained granite. Numerous

quarries near Llano produce a fine-grained gray monumental granite—the Sixmile granites (Sellards and Baker, 1934, p. 75). This gray granite cuts coarse porphyritic granite at the Curley quarry, half way between Llano and Valley Spring, in the same manner as the fine-grained granite cuts the outer rim of coarser granites in the massifs studied. It would seem, therefore, that the fine-grained granites in this region are younger than the coarse-grained varieties and probably represent later phases of the same period of intrusion.

The llanite or "opaline" quartz porphyry (Iddings, 1904) discussed in connection with the granites of the Buchanan massif cuts all other pre-Cambrian rocks and is one of the last phases of the granite intrusion. It is the only true porphyry in the region.

DISCUSSION OF THE MODE OF INTRUSION

GENERAL FEATURES OF MASSIFS

In considering the mode of intrusion several features of the massifs must be taken into consideration: (1) The granite varies texturally with the finer-grained varieties in the centers of the massifs; (2) the contacts between bands of textural types are in the typical case gradational; (3) roof pendants or screens of country rock separating different phases of the intrusion are lacking; (4) the porphyritic type of granite occurs typically not in the outermost or earliest crystallized band, but in an intermediate position; and (5) the contact with the wall rock is always sharp.

COMPARISON WITH OTHER REGIONS

Occurrences of porphyritic and nonporphyritic granite from other regions are similar in some respects to the Texas granites. At Philipsburg, Montana, Emmons and Calkins (1913, p. 115-117) describe a porphyritic biotite granite surrounded by an earlier nonporphyritic biotite granite. The sequence of intrusion of the granites at Dartmoor, England, is summarized by Osman (1928, p. 259-261) as (1) rafts of basic microgranite, (2) coarse groundmass crowded with feldspar phenocrysts (up to 7 inches long and exhibiting platy parallelism), (3) stages where the phenocrysts are not crowded and the groundmass is finer, (4) marginal intrusions of fine-grained granite, and (5) aplite, pegmatite, quartz veins, and felsite dikes. Many of the granites of the southeastern Atlantic States described by Watson (1910) seem to be similar to the Texas occurrences, although there are no exact parallels. The coarse-grained pink granite quarried at Stony Creek, Connecticut (Dale and Gregory, 1911, p. 79-84) has been interchanged with the Texas granite by archi-

tects. The Connecticut granite, however, is richer in potash, has more xenoliths and roof pendants, more pronounced flow structure, and does not show the same textural variation.

Further illustrations could be given of granites similar to those described in this paper, but the writer knows of no described examples which show just the same combination of features as are found in the Llano-Burnet region.

ALTERNATE THEORIES OF INTRUSION

Theories of intrusion proposed for igneous bodies elsewhere are not considered to be valid for the massifs described in this paper for the following reasons:

(1) The possibility that the concentric structure is due to granitization of the country rock as described by Anderson (1937) in the Inyo Mountains is ruled out by the sharp contacts with the wall rock (Fig. 4) and the lack of any transitional phases between schist and granite. (2) Ring-dikes produced by the subsidence of blocks of country rock as described from Great Britain (Richey, 1932) and New Hampshire (Modell, 1936; Kingsley, 1931) differ in several respects from the massifs described in this paper. They are generally smaller in area, there is a greater diversity in the rock types, the contacts between separate phases are sharp, and there are portions of the country rock included within the intrusion. (3) There is no evidence that stoping was a major factor in the process of intrusion. Xenoliths and roof pendants are present only around the margins where they are parallel to flow layers as if carried upward with the intruding magma. The blocks of earlier porphyritic granite described in the Kingsland and Katemcy massifs may be completely altered portions of wall rock, but there is no evidence that they have fallen rather than risen.

PROPOSED THEORY OF INTRUSION

General statement.—The massifs of the Llano-Burnet region are thought to have originated by forcible injection rather than by any of the processes outlined above. Figure 12 is designed to show earlier stages in the intrusion as pictured by the writer. The magma is thought to have intruded vertically into the crust. If granitization and assimilation of wall rock was a factor, the process must have taken place at greater depths, and the products must have been thoroughly incorporated into the magma which later rose and cut with sharp contacts past higher schist and gneiss. The restored upper portion of the intrusion is shown as a dome, though the upward extension may have been a series of dikes or some other form open to the surface (Balk, 1937, p. 87). As the magma

rose it also expanded laterally, so that movement in the outer portions was about the same in all directions, as is borne out by the lack of linear structures around the margins.

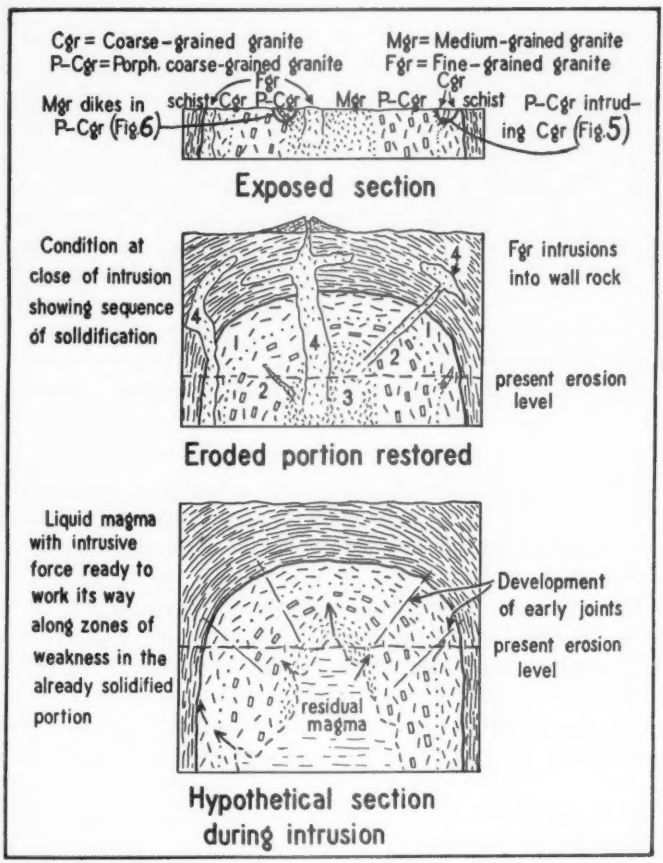

FIGURE 12.—*Sections showing suggested earlier stages of typical massif*
Width of sections, about 12 miles. Vertical scale hypothetical.

The conditions shown in the hypothetical sections at the bottom of Figure 12 are those postulated before the central part of the intrusion had crystallized. Zones of weakness, such as developing joints or the contact zones, are indicated as the location of subsequent intrusions of

fine-grained granite. It is suggested that they are derived from the residual portions of the magma because of their petrographic similarity to the coarser granites. There is nothing in the field evidence, however, to show that they may not have been intruded at a much later date.

Origin of porphyritic granite.—In the case of porphyritic rocks it is of interest to know whether phenocrysts are intratelluric or have developed *in situ* (Watson, 1901). The reasons why the phenocrysts are believed to be primarily intratelluric have been mentioned in the discussion of the porphyritic coarse-grained granite of the Buchanan massif, but the development of porphyritic coarse-grained granite after coarse-grained granite is difficult to explain. The former is transitional to the latter and in some cases is in contact with the wall rock. Presumably the large microcline and microperthite phenocrysts were growing all the time but did not become large or numerous enough to give a porphyritic appearance until crystallization had been in progress some time. After they had accumulated in considerable numbers, the phenocrysts may have been further concentrated by a process of pushing aside by the magma still being intruded into the center of the chamber.

Origin of finer-grained central core.—The reason for the smaller grain size of the core of medium-grained granite is not clear. Conditions for the central part of the magma chamber must have changed, causing the average grain size to decrease (although still maintaining some of the larger phenocrysts locally). This is contrary to the normal situation where the central part of an intrusion is coarser-grained than the margins. The following suggestions are offered as working hypotheses.

(1) The rate of movement of the magma may have increased as the outer portions of the massif crystallized and reduced the size of the conduit (assuming that the massif had a connection with the surface) thus increasing the rate of crystallization, as a result of greater movement within the liquid, and decreasing the average grain size. In the Buchanan massif the contact between the porphyritic coarse-grained granite and the medium-grained granite is abrupt, indicating a renewed surge in the intrusion of the latter into the former. Usually, however, there is no evidence of greater motion in the cores of the massifs, and the rock in the central portions is often devoid of flow structures.

(2) Increase in the rate of cooling may be responsible for the finer grain size of the centers, although by what mechanism the rate of cooling may have increased it is not possible to say. As pointed out by Lovering (1935) the heat conductivity of rocks is exceedingly slow.

If the time required for the solidification of a batholith 15 miles in diameter were long enough for considerable erosion to have taken place from the top of the dome produced by the intrusion of the batholith, the central portion of the magma chamber may have been closer to the surface when it crystallized than the outer portions had been, and the rate of heat loss would consequently have been more rapid. It seems improbable that the rate of erosion would have been rapid enough for this, but in the lack of definite data the suggestion is offered as a possibility.

(3) The progressive concentration of volatiles in the residual magma in the center of the chamber would be expected to increase rather than diminish the size of the crystals. But if the magma chamber (which, up to the time of crystallization of the central portion, had not been connected with the surface) suddenly became the source of extrusive activity, it is possible that the residual material left in the center of the chamber would have become impoverished in volatiles as a result of the extrusion of the volatile-rich material. Thus the core of the massif could have crystallized with a smaller proportion of volatiles present than in the outer zones and thus have developed a smaller grain size.

SUMMARY AND CONCLUSIONS

(1) The massifs are steep-walled, roughly circular bodies, 10 to 12 miles in diameter, flanked by schist and gneiss.

(2) Within the massifs there are concentric bands of different textural varieties of granite—coarse-grained granite on the outside, porphyritic coarse-grained granite in the intermediate ring, and medium-grained granite (often containing occasional phenocrysts) in the core.

(3) Contacts between the rocks of different texture are usually gradational.

(4) The concentric bands are cut by small bodies of fine-grained granite in the center of the massifs, around the margins, and in radial dikes and elongated masses.

(5) The sequence of solidification for the textural varieties is shown to be as follows: coarse-grained granite, porphyritic coarse-grained granite, medium-grained granite, and fine-grained granite.

(6) Comparison of the type massif of the area with other granite occurrences reveals a repetition of the same general features, although there are exceptions. The most common textural type for the region as a whole is porphyritic coarse-grained granite.

(7) It is believed that the massifs were intruded vertically into the crust and that differentiation into textural phases accompanied the

intrusion. Suggestions are offered to explain the porphyritic texture and the finer grain size of the central core as compared with the margins.

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EXPERIMENTAL FLOW OF ROCKS UNDER CONDITIONS FAVORING RECRYSTALLIZATION[1]

BY DAVID GRIGGS

CONTENTS

	Page
Abstract	1002
Introduction	1002
Relative importance of the different mechanisms of flow	1003
Failure of high pressure alone to duplicate natural deformation	1005
Experiments on limestone and marble	1005
Quartz experiments	1007
Effect of time alone	1008
Creep of alabaster in the presence of solutions	1008
General statement	1008
Effect of varying solutions	1009
Effect of varying stress	1013
Delayed fracture	1017
Strength	1018
Tests under moderate confining pressure	1020
Conclusion	1021
Works to which reference is made	1022

ILLUSTRATIONS

Figure	Page
1. Fabric diagram of Yule marble before deformation	1006
2. Fabric diagram of Yule marble after deformation	1006
3. Stress-strain diagram of Ohio alabaster	1010
4. Elastic flow of alabaster in normal creep test	1011
5. Creep of alabaster in different chemical environments	1012
6. Assembled creep curves of alabaster	1013
7. Creep rate of alabaster as a function of stress	1014
8. Creep of alabaster at low stress	1015
9. Creep rate of Solenhofen as a function of stress	1016
10. Strength of alabaster as a function of duration of test	1019

Plate	Facing page
1. Bank of creep testers and Solenhofen limestone	1004
2. Photomicrographs of Yule marble	1005

[1] After these experiments had been largely completed, the writer learned that some years previously Dr. Roy W. Goranson had made a thermodynamic analysis of recrystallization flow in stressed solids, and that the relations so derived correspond closely to some of the empirical relations found in the experiments. This previously unpublished work appears in the following paper, chronologic development being generously sacrificed by Dr. Goranson to promote continuity in the presentation.

ABSTRACT

In order that the results of experimental investigations of the flow of rocks may be applied with any degree of assurance to problems of rock deformation in nature, it is essential that the effects of all five of the varying environmental factors be investigated: (1) confining pressure, (2) shear stress, (3) temperature, (4) time, and (5) the presence of solutions. Pressure alone changes the behavior of rocks and simulates some types of natural deformation but is shown to be inadequate to explain all natural flow because of the tremendous strength increase involved, its inability to produce plastic behavior in quartz, and the development of too intense mechanical twinning in calcite aggregates. Temperature produced unimportant results in preliminary experiments at high pressure. Time seems to be inadequate as shown by creep tests on Solenhofen limestone at low and high pressure. Some new experiments show the effect of solutions. The properties of quartz change markedly in the presence of solutions at elevated temperatures. Marble flow is strikingly altered by the presence of solutions at elevated temperature. New creep tests on alabaster in the presence of solutions seem to demonstrate recrystallization flow and indicate the existence of two general laws governing this type of flow. These reconnaissance experiments indicate the important role that solution and recrystallization may play in rock deformation.

INTRODUCTION

"Flow" is often a controversial term because of its many connotations in geological literature. In laboratory experimentation, as in the field, it is often impossible to observe enough criteria to define completely the deformation of rocks. In this paper it will be used in the broadest sense.

The behavior of a body under stress may be divided into three parts: (1) elastic deformation—practically instantaneously achieved and as rapidly reversible on removal of the external forces acting on the body; (2) flow—that part of the deformation in excess of the elastic deformation but which does not result in a notable loss of cohesion. It is characterized by change of shape as a function of time, at constant stress; (3) rupture—that part of the deformation characterized by loss of cohesion. Frequently flow grades into rupture, with a progressive loss of cohesion until complete separation occurs.

Flow is commonly classified under the two divisions—viscous flow and plastic flow. The laws of viscous flow are simple, have been well defined by physicists, and are incorporated into the general treatment of hydrodynamics. "Plastic flow" on the other hand is a term used to cover a multitude of different behaviors, some of which are imperfectly understood. It seems that no useful end will be served by the use of the term in the present discussion, so it will be avoided. No viscous flow has been detected in the crystalline aggregates tested to date in the laboratory. The flow that has been observed will be designated by descriptive terms which may need revision as our understanding of the mechanism progresses.

RELATIVE IMPORTANCE OF THE DIFFERENT MECHANISMS OF FLOW

The materials with which this paper is concerned are the compact crystalline rocks and not loosely consolidated sediments—crystalline aggregates in constitution, not weakly cemented discrete particles. The mechanisms of flow in crystalline aggregates may be classified in three general categories:

1. Cataclasis—deformation characterized by mechanical granulation.
2. Gliding—intragranular twin- or translation-gliding dominant, with secondary intergranular adjustment consequent on the change of grain shape resulting from intragranular motion.
3. Recrystallization—deformation by molecular rearrangement through the medium of solutions, local melting, or solid diffusion; intergranular motion dominant.

The relative importance of each of the three mechanisms of flow has been the subject of violent controversy for decades. The evidence presented for each viewpoint has been largely that of detailed study of rock fabrics—first megascopically, then with simple microanalysis, and more recently with the revolutionary statistical microscopic methods of Sander and Schmidt.

The controversy about mechanism of flow has hinged on the question of whether mineral orientation in rocks is developed by shearing, which aligns grains and glide-planes in the shear planes, or by recrystallization, forming flow cleavage by the growth of crystals perpendicular to the direction of maximum compression. In partial reconciliation of the various viewpoints, the concept of zones of fracture and flow was adopted, with the result that one could insure his preferred mechanism by proper choice of the hypothetical environment in which the deformation was supposed to have occurred.

Petrofabric analysis has brought a wealth of objective data bearing on the problem. Interpretation has lagged far behind this accumulation of data, however, for the reason that the mechanism of rock deformation is largely unknown. In the last analysis, every interpretation of a fabric diagram depends upon hypotheses of mineral orientation in part analytical and in part developed from analogies with the fabrics of deformed metal aggregates.

Insofar as interpretation depends on observed geologic structures, it may be said conservatively that the field data are not wholly satisfactory. A large portion of the most detailed petrofabric analysis has been done in the Eastern Alps, where, to say the least, the major structure is uncertain. It is to be hoped that detailed analyses will be made in areas in which a simple geologic structure is well known in three dimensions—as, for example, in highly developed mining camps—to show the relation between the mineral orientations developed and the structure.

The analytical part of the interpretation rests largely on the assumption that rock cleavage is the result of shearing and that some crystal vectors will be oriented in the shear planes. It is possible to get a considerable amount of internal evidence by comparison of the fabrics of the various mineral constituents of the rock, and this method has been used to derive several hypothetical mechanisms of crystal deformation—twin- and translation-glide planes which have yet to be demonstrated in the laboratory. Because of the complexity of the structures and the geological history of the areas in which these studies have been made, all these interpretations rest on assumptions which are difficult to verify.

Insofar as the interpretation of fabrics is based on analogy with the behavior of crystalline metal aggregates, one must consider the justification of such analogies. Rock deformation superficially seems to have much in common with the deformation of metals. They both exhibit twin- and translation-gliding; they both deform by recrystallization. Other analogous physical properties are work-hardening, elastic flow, and creep. On the other hand, the atomic bonding of the metal crystals is very different from that of the rock-forming minerals, and the rules relating atomic structure to deformation developed for some metals do not hold for rock minerals. Metals readily recrystallize in the dry state when a certain critical temperature is exceeded, but similar recrystallization has not been produced on an observable scale in the rock-forming minerals with which the writer has experimented. It seems necessary that the latter be in contact with solutions.

It is to be hoped that laboratory experimentation on the deformation of rocks will show how far the analogies with metal behavior are applicable to the interpretation of rock fabrics and in what respects it is necessary to develop new laws of flow. The experiments described in the present paper strongly indicate that, unlike the deformation of metals, solution and recrystallization play an important role in rock deformation. Goranson suggests, however, that similar thermodynamic equations describe the dry flow of metals and the recrystallization flow of rocks.

It has been observed in metals that the conditions of testing produce markedly different mechanisms of deformation. For example, a metal deformed rapidly at low temperature usually flows by intracrystalline gliding. On the other hand, the same metal tested at high temperature or deformed very slowly at low temperature may flow primarily by intergranular motion which seems connected with recrystallization or solid diffusion, and gliding may be secondary or even entirely absent (Moore, Betty, and Dollins, 1935). Similarly, it is of the greatest importance in laboratory experiments on the deformation of rocks to observe the effects of all five of the variable environmental factors in conditions of

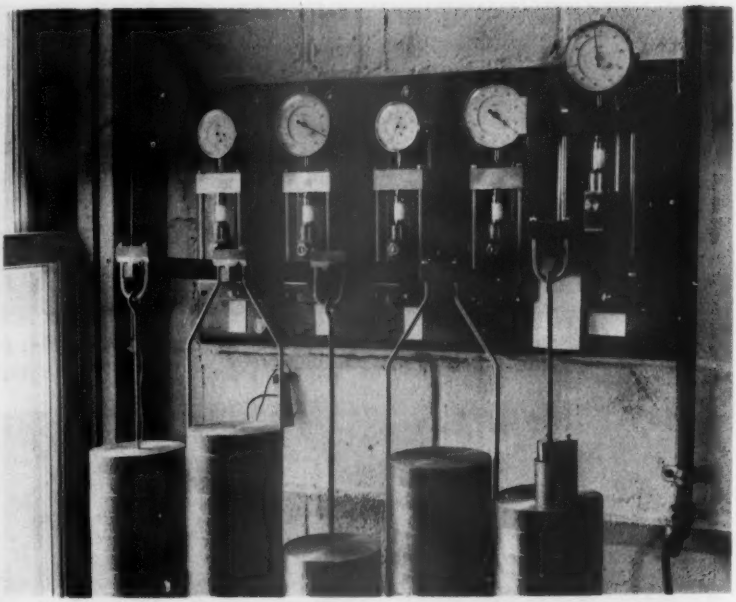

FIGURE 1. BANK OF CREEP TESTERS
With specimens of alabaster in place. Compressive stress applied by weights on lever arms; specimens surrounded by small jacket of water in glass tubes; creep measured by dial gauges.

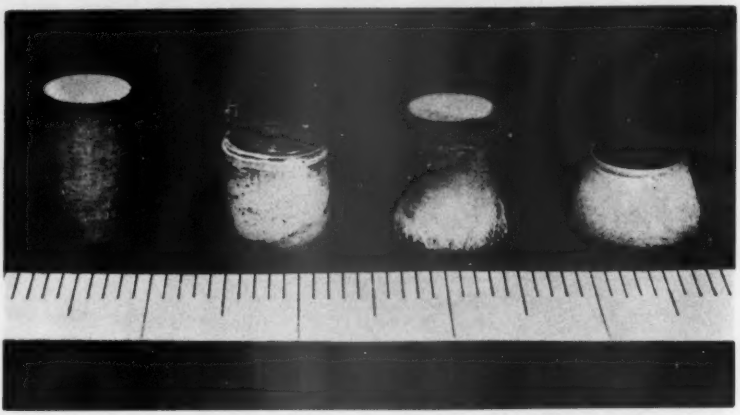

FIGURE 2. SOLENHOFEN LIMESTONE
Deformed under high pressure, undeformed specimen on left for comparison; second from left shows relatively uniform flow under pressure of about 10,000 atmospheres; next from left shows incipient fracturing and nonuniform deformation at about 2000 atmospheres; specimen on right shortened 45 per cent at about 7000 atmospheres.

BANK OF CREEP TESTERS AND SOLENHOFEN LIMESTONE

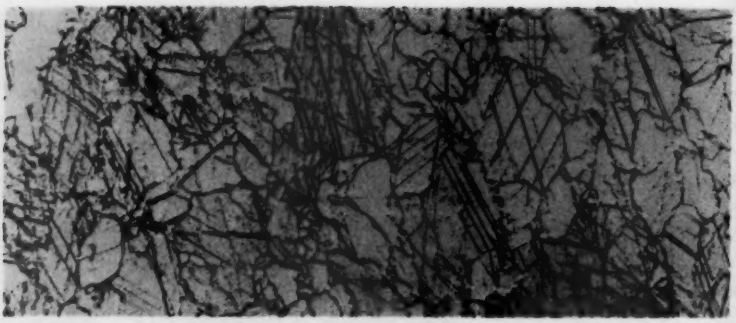

FIGURE 1. BEFORE DEFORMATION
25x, plane polarized light.

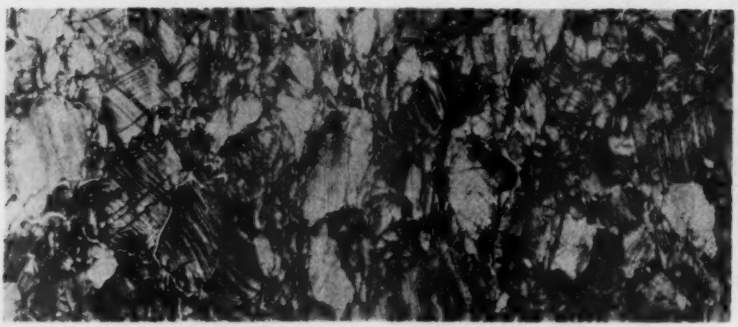

FIGURE 2. AFTER 30 PER CENT SHORTENING
At 10,000 atmospheres confining pressure, 150° C

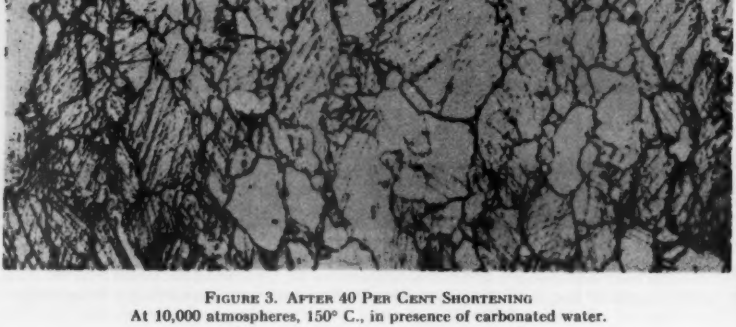

FIGURE 3. AFTER 40 PER CENT SHORTENING
At 10,000 atmospheres, 150° C., in presence of carbonated water.

PHOTOMICROGRAPHS OF YULE MARBLE

geological deformation: confining pressure, shear stress, temperature, time, and the presence of solutions. Only when the effects of all these variables have been evaluated in the laboratory can we arrive at an understanding of the mechanism of deformation of rocks in nature.

This paper shows some steps in successive approximation of the effects of the five variables separately and some combinations of the variables. It is important to note the nature of these variables. Two are truly independent variables: confining pressure and temperature. Chemical environment, in the form of solutions, is dependent on both of these two in a closed system. Shear stress is dependent on the ability of the rock to sustain it, and this ability is a function of all three of the above variables and also of time. The duration of a test and the shear stress are mutually dependent, as shown in some of the experiments to be described.

FAILURE OF HIGH PRESSURE ALONE TO DUPLICATE NATURAL DEFORMATION

EXPERIMENTS ON LIMESTONE AND MARBLE

Various experiments, such as those of Adams, von Karman, and the writer, have shown that many rocks which are brittle in our ordinary experience change their properties under high confining pressure and possess a high degree of plasticity. It is our purpose here to review this pressure-induced plasticity in order to show in what measure the result is similar to natural deformation and how the dissimilarities suggest the inadequacy of high pressure alone in explaining the physics of rock deformation.

Solenhofen limestone under ordinary conditions exhibits only elastic behavior. No permanent set, or flow, is observed. When surrounded by a hydrostatic pressure of 10,000 atmospheres, however, it may be shortened 50 per cent without fracture by a suitable compressive stress. Figure 2 of Plate 1 shows three specimens of Solenhofen limestone which have been considerably deformed under varying pressure conditions in a simple adaptation of Professor P. W. Bridgman's high pressure apparatus (Griggs and Bell, 1938, p. 1730).

Concomitant with this increase in malleability is a great increase in strength of the material. Solenhofen limestone subjected to a liquid confining pressure equivalent to that at a depth of 22 miles in the earth's crust exhibits an increase in strength in short-time tests from its normal 2600 kg/cm^2 to 13,000 kg/cm^2, and extrapolation of the strength-confining pressure curve would indicate an even greater increase in strength with additional confining pressure. It seems unlikely that a differential stress great enough to overcome this strength can be developed within the

crust of the earth. Geological observations indicate that the earth's crust deforms under differential stresses of the order of magnitude of 1000 kg/cm². This implies that, although pressure alone may produce flow of limestone and marble, other factors must operate to reduce the strength

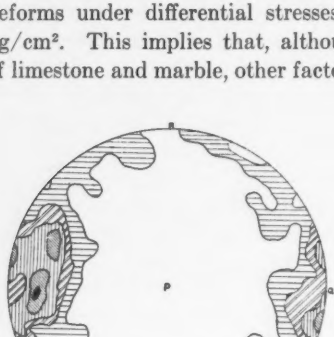

FIGURE 1.—*Fabric diagram of Yule marble before deformation*

Directions P, Q, R are the geographic coordinates: vertical, E-W, and N-S, respectively. Shows high degree of preferred orientation in original marble. 209 calcite axes, contours 9, 7, 5, 3, 1 per cent. Measured by Dr. E. B. Knopf.

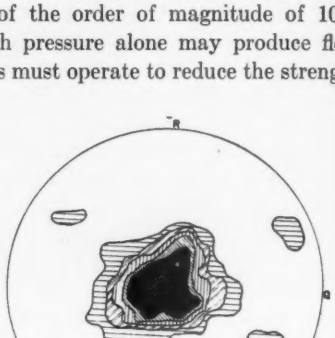

FIGURE 2.—*Fabric diagram of Yule marble after deformation*

Compressed 30 per cent parallel to P fabric axis, or parallel to the resultant axes maximum, 10,000 atmospheres confining pressure. 163 calcite axes, contours same as figure 1. Measured by Dr. J. F. Bell.

of the materials under the high confining pressures existing in the earth's crust.

Experiments on calcite single crystals show that mechanical twinning on the $10\bar{1}2$ plane is the most important mechanism of deformation under confining pressures up to 10,000 atmospheres (Griggs, 1938). Deformation of marble under pressure alone, and under pressure and elevated temperature, appears to occur primarily by twinning. Figures 1 and 2 of Plate 2 show the change in texture of Yule marble when shortened 30 per cent at a confining pressure of 10,000 atmospheres and 150° C. These experiments were performed in co-operation with Dr. E. B. Knopf, who has made detailed measurements of the marble fabric before and after deformation. Although it is too early to present a complete report of the results, petrofabric diagrams from the sections illustrated in Figures 1 and 2 of Plate 2 are shown in text Figures 1 and 2.

Analyzed thin sections from several parts of the marble block used showed a relatively homogeneous fabric, so that we can be sure that the fabric of the specimen shown in Figure 2 was essentially that of

Figure 1 before deformation. It is seen that the 30 per cent shortening under high confining pressure completely altered the fabric, obliterating the old maximum and developing a high degree of optic axis orientation parallel to the direction of compression. The twinning in this specimen is more intense than that in many natural marbles which we know have been more drastically deformed.

This preliminary experiment on the deformation of marble under elevated temperature showed no significant change in behavior from that at room temperature, other conditions being the same. Another experiment under the same conditions of temperature and pressure, but with carbonated water surrounding the marble specimen, showed a great difference in behavior, however. The stress necessary for 30 per cent shortening was only 1500 kg/cm^2, compared to 8100 in the former experiment, and the texture of the deformed marble was entirely different (Pl. 2, fig. 3). Here microscopic study of the calcite showed very slight development of twinning and in fact showed little change in the texture of the calcite from that of the original, even though the plastic flow of the specimen was greater than that shown in Figure 2 of Plate 2 (40 per cent shortening). It has been impossible to determine by petrofabric analysis the mechanism of flow in this last experiment because the specimen was unfortunately compressed parallel to the pre-existent direction of preferred orientation of axes, and the fabric was not greatly changed. The presence of water and CO_2 would favor solution and recrystallization of the individual calcite grains, and deformation might have proceeded by recrystallization flow.

QUARTZ EXPERIMENTS

It is a familiar geological observation that quartz is mobile in rocks under metamorphic conditions. Experiments in which crystals of quartz were subjected to the highest pressure then available in laboratory experimentation failed to produce any observable flow of the quartz (Griggs and Bell, 1938, p. 1730-1732). In these experiments, as with the Solenhofen limestone, the strength of the quartz increased very greatly with an increase in confining pressure. Under a confining pressure of 19,500 atmospheres the strength of the quartz was 118,500 kg/cm^2. Bridgman (1940) has recently extended this to 25,000 atmospheres confining pressure and has found quartz still brittle, with a strength of 150,000 kg/cm^2. Even under this colossal differential stress the quartz exhibited no flow but only elastic behavior up to the point at which fracture occurred. These experiments were performed at room temperature, and the quartz was dry. In another series of experiments (Griggs and Bell, 1938, p. 1733-1736) quartz single crystals were deformed at temperatures up to

400° C. when surrounded by aqueous solutions which provided a few hundred atmospheres of confining pressure. In these experiments, as in the former series, no flow was observed, but the strength of the quartz was materially lowered by the combination of high temperature and solutions. A differential stress of 4000 kg/cm² was sufficient to fracture the quartz after several hours under the influence of high temperature and solutions. The nature of the fracture in the latter experiments was very different from that in the former, and the hypothesis was advanced that the orientation of quartz in deformed rocks might occur as a result of comminution of this type and orientation of the developed needles in the shearing accompanying rock deformation.

New apparatus permits more complete observation of the behavior of quartz at high temperatures and in the presence of solutions. Although it is too early to report definitive results of these experiments, it seems evident that quartz becomes mobile under conditions favoring recrystallization flow in a way that seems impossible under other conditions.

EFFECT OF TIME ALONE

A recent paper (Griggs, 1939) presented the results of some experiments on creep of rocks, or the slow deformation under small stresses acting over long periods of time. With the exception of one preliminary test on recrystallization flow of alabaster, these tests were all performed on dry materials in conditions unfavorable to solution and recrystallization. The crystalline aggregates tested all exhibited elasticoviscous flow in which elastic flow predominated and pseudoviscous flow was small or absent. Solenhofen limestone, for example, subject to a compressive stress of 1400 kg/cm² for a long period, showed purely elastic flow of an inconsequential magnitude compared to behavior inferred from geology. This experiment has now been in operation for a period of 900 days and shows no significant deviation from this elastic flow.

The following section shows that alabaster when dry exhibits the same type of flow as the Solenhofen, but its behavior in the presence of solutions is entirely different.

CREEP OF ALABASTER IN THE PRESENCE OF SOLUTIONS

GENERAL STATEMENT

Gypsum was chosen as a rock-forming mineral which would be susceptible to recrystallization under conditions easily attainable in laboratory creep tests. Because of its relatively high solubility in water at room temperature and atmopheric pressure (one part in five hundred) it was thought likely that solution and recrystallization would occur under

constantly applied stress in accordance with Riecke's principle and thus produce recrystallization flow.

Riecke (1895) derived thermodynamic equations for the lowering of the melting point by the action of differential stress, and these apply equally well to the increase in solubility. Russell (1935) showed experimentally that stresses may be localized within a single crystal and that, because of this, solution and recrystallization may take place on one crystal grain under localized stress, with consequent flattening normal to the direction of compression. In his experiments the crystal was in contact with its own saturated solution.

From such considerations, it might be expected that a crystalline aggregate in contact with its own saturated solution and subject to differential stress would have a tendency to deform by solution at points of stress concentration and recrystallization at points where the stress is lower. It might be further expected that grain elongation would occur normal to the direction of maximum compression. In addition, due to different solubilities of crystals in different crystallographic directions, it might be expected that crystals with certain orientations would be more rapidly dissolved and other orientations might be preserved and enlarged at their expense, thus providing a mechanism for the development of mineral orientation in deformed rocks.

These expectations have been verified to only a small extent by the present experiments, but it is believed that conditions of recrystallization flow have been attained, and the resulting effects seem suggestive enough to warrant description in detail.

EFFECT OF VARYING SOLUTIONS

In a new apparatus (Pl. 1, fig. 1) financed by a project grant from the Penrose Bequest of the Geological Society of America the writer has performed some creep experiments on specimens of alabaster in contact with solutions. A preliminary experiment of this type was reported previously (Griggs, 1939, p. 247). The new experiments deal with specimens from a different piece of alabaster which is finer in grain and stronger than that used in the preliminary test. All the experiments have been performed at room temperature. In each test the differential stress (compressive stress) on the specimen has been constant, but various stress differences have been used in the different experiments to determine the effect of varying stress on the properties of flow. In addition to variation in magnitude of stress from experiment to experiment, the specimens have been surrounded by different solutions, so as to observe the effect produced in this way. Finally, the effect of confining pressure has been investigated.

The behavior of the alabaster in a normal short-time compressive test is shown in Figure 3. The deformation is primarily elastic, and there is very slight flow before fracture occurs. The strength is in excess of 500 kg/cm², and the shortening before rupture about .3 per cent. A specimen

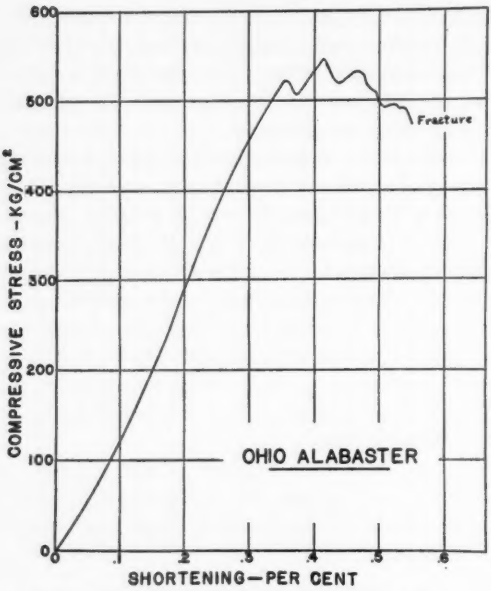

FIGURE 3.—*Stress-strain diagram of Ohio alabaster*
In short-time compressive test, 24° C., no solutions.

subjected to a differential stress of 420 kg/cm² for a period of 40 days exhibited purely elastic flow, as shown in Figure 4, and the total amount of this elastic flow under stress was only .07 per cent.

An entirely different type of behavior is observed when the alabaster is surrounded by its own saturated solution. Figure 5 shows the behavior of three alabaster specimens under the same value of differential stress (205 kg/cm²) and at the same temperature, but in the presence of different solutions. Creep in the dry state (the bottom curve in Figure 5) is seen to be a small amount of elastic flow, but, when the specimen is surrounded by a small jacket of distilled water which soon becomes saturated with gypsum, the creep rate is greatly accelerated and exhibits an entirely different behavior. This creep curve in the presence of water may be

divided into three stages of creep similar to the three stages observed in the creep of some metals (Moore, Betty, and Dollins, 1935): (1) creep at a decreasing rate, (2) creep at a constant rate (pseudo-viscous flow), and (3) creep at an increasing rate, leading to fracture.

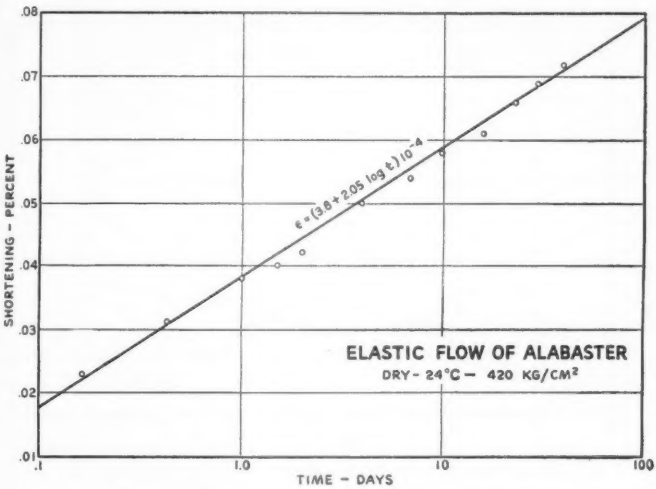

FIGURE 4.—*Elastic flow of alabaster in normal creep test*
Under load of 420 kg/cm².

It was suggested to the writer that the increased rate of creep in the presence of water might be caused by the water acting as an interstitial lubricant. To test this a creep test was run, in which the specimen was surrounded by a jacket of Tasgon penetrating oil. Although this oil appeared to fill the pores of the rock, the observed creep was no different from that in the dry state.

If the deformation in water were due to flow by solution and recrystallization of the constituent grains—solution at points of greatest stress and deposition at points of lesser stress—it might be expected that creep would be further accelerated if the specimen were surrounded by a solution in which the solubility of gypsum is higher than in water. To investigate this, a creep test was run under the same conditions as before, except that the alabaster was surrounded by a dilute HCl solution. As shown in Figure 5, creep under these conditions was considerably more rapid than in the presence of water.

Dr. E. B. Wilson, Jr., suggested that if the rate of creep were proportional to the solubility as indicated by this experiment, then creep would

be slower in an aqueous solution in which the solubility was lower than in distilled water. For this purpose the common ion effect of calcium chloride was employed. A creep test was run under similar conditions of stress and temperature, but the specimen was surrounded by a solution

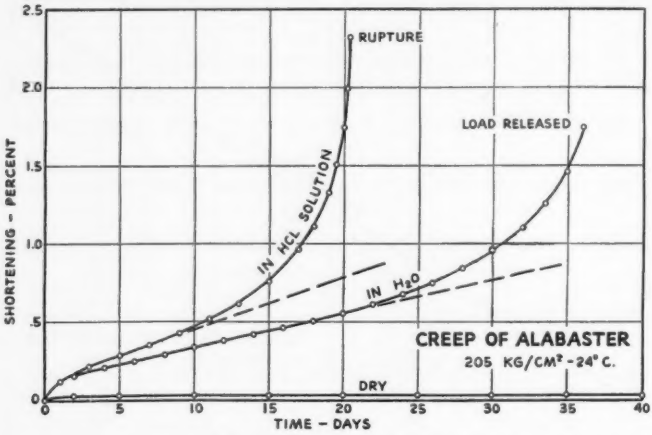

FIGURE 5.—*Creep of alabaster in different chemical environments*
Three specimens, all loaded to 205 kg/cm², all at the same temperature, but one dry, one in distilled water, and one in dilute HCl.

of calcium chloride (16.9 gr/l) in which the solubility of gypsum was lowered approximately 45 per cent. Contrary to expectations, creep under these conditions was more rapid than in pure water. From this it may be inferred that the rate of creep is not a function of the actual amount of the gypsum that may be present in the solvent—*i.e.*, the solubility—but may be a function of the ionic mobility of the gypsum in the solvent (Goranson, 1940, p. 1027).

Microscopic analysis of thin sections has provided no further clues to the mechanism of this deformation. As will be noticed in Figure 5, the deformation is small—less than 4 per cent—and this slight amount of flow has not produced changes in the crystalline structure of the alabaster great enough to be observed with ordinary microscopic technique.

One might think that penetration of the water into the interstices of the alabaster would take sufficient time so that this penetration would have a large effect on the nature of the creep curve. In the experiments of Figure 5 the stress was applied immediately after the specimen had been surrounded with water. A similar series of experiments has been

performed in which the alabaster was soaked for a period of 68 days in distilled water, and then the stress applied. The soaked alabaster exhibits the same behavior as that in which there was no previous soaking, within the limits of reproducibility of the specimens.

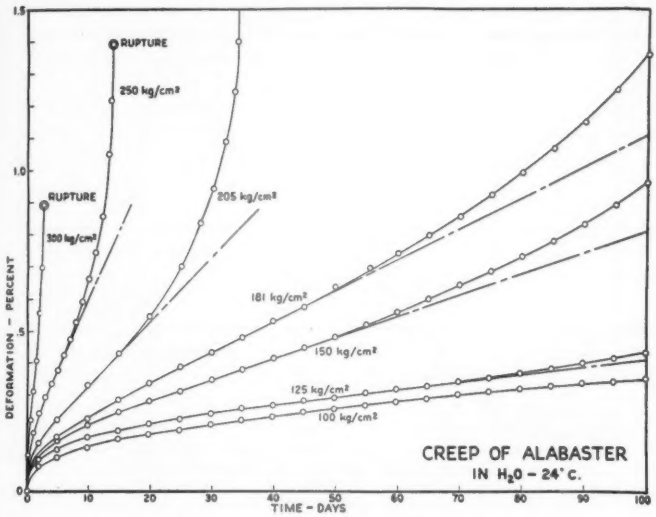

FIGURE 6.—*Assembled creep curves of alabaster*
Seven specimens, all in distilled water, at the same temperature, but loaded differently.

EFFECT OF VARYING STRESS

Figure 5 shows a characteristic curve of creep in alabaster at room temperature, surrounded by a jacket of water. It is found that deformation of this type occurs under varying values of differential stress. Figure 6 is a collection of a series of seven creep tests under different values of stress. Although the normal strength of the alabaster is greater than 500 kg/cm², we see here that when immersed in water and subjected to a differential stress of 300 kg/cm² it flows very rapidly, and rupture occurs in less than 3 days of constant application of this stress difference. (Compare with Figure 4.) Under a stress of 250 kg/cm² flow proceeds at a slower rate and rupture does not occur until almost 14 days have elapsed, and, in addition, the deformation before rupture is much greater than at 300 kg/cm². Under stresses of 225, 205, 181, 165, 150, and 125 kg/cm² similar behavior was observed in that after creep exceeded a certain amount the rate increased, leading to fracture. The behavior

of the specimen at 103 kg/cm² was somewhat different and will receive attention later.

If we consider that the part of the creep in which the rate is constant represents an equilibrium state of recrystallization flow, then we may

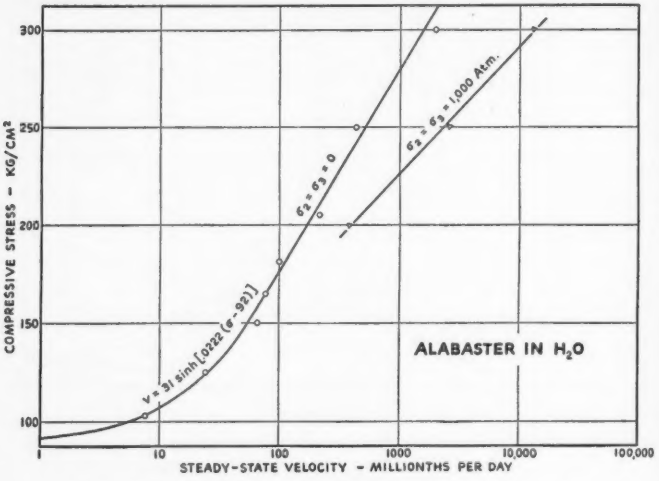

FIGURE 7.—*Creep rate of alabaster as a function of stress*

Upper curve from experiments of Figure 6; Lower curve from creep tests under 1000 atmosphere confining pressure.

take this as a fundamental constant of creep. Plotting this steady-state velocity against the stress, we find the relation shown in Figure 7, in which, to a first approximation:

$$v = a \sinh b (\sigma - S) \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \quad (1)$$

(v = steady-state velocity, σ = compressive stress, a, b, and S are empirical constants)

It is seen that as 100 kg/cm² is approached, the creep rate decreases markedly, and, if equation (1) is correct, this rate will become zero at a stress of 92 kg/cm². This "fundamental strength" of the alabaster under the given conditions (Griggs, 1936, p. 564) is our constant "S" in equation (1).

Since this concept of fundamental strength is one of considerable importance to geology and since this is the first quantitative indication of such a strength under conditions of recrystallization flow, it is important to examine the experiments closely for some further clue as to mechanism. The test at 103 kg/cm² provides some suggestive information. Figure 8

shows the full curve. For the first 300 days the flow can be approximated by a power function, so that the strain-time relation is linear when plotted on log-log paper. Since this is not true of the tests at higher stress, it may be indicative of a different mechanism of flow.

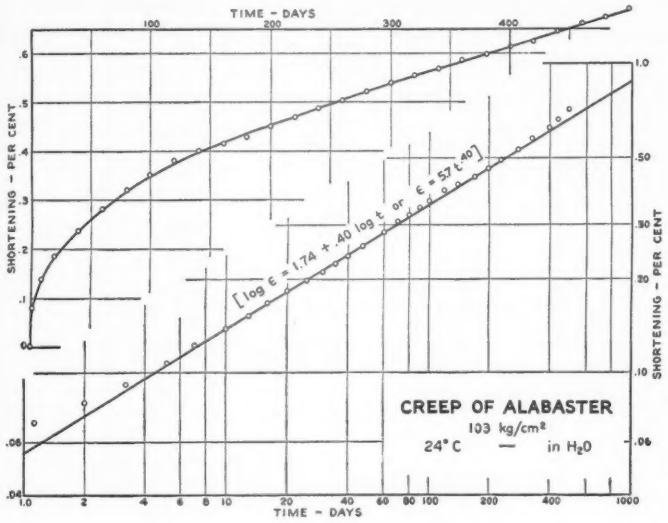
TIME - DAYS
SHORTENING - PER CENT
[log ε = 1.74 + .40 log t or ε = 57 t^.40]
CREEP OF ALABASTER
103 kg/cm²
24° C — in H₂0
TIME - DAYS
SHORTENING - PER CENT

FIGURE 8.—*Creep of alabaster at low stress*

Upper curve: plotted on normal co-ordinates; lower: same data plotted on double logarithmic co-ordinates.

If we write equation (1) in exponential form,

$$v = \frac{a}{2}\left[e^{b(\sigma - S)} - e^{-b(\sigma - S)}\right] \quad \ldots\ldots\ldots\ldots\ldots\ldots (2)$$

we see that when σ is considerably in excess of the fundamental strength, the second term becomes negligible, and:

$$b = \frac{a}{2} e^{b(\sigma - S)} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (3)$$

so that the σ — v curve is a straight line when plotted on semi-logarithmic co-ordinates.

This information on the dependence of flow rate on stress is helpful in testing hypothetical mechanisms of flow. The nature of the environment immediately suggests flow by solution and recrystallization. In the following paper, Goranson derives from thermodynamic considerations equations of recrystallization flow which are exactly similar in form to

these empirical equations. This, coupled with the fact that his thermodynamic treatment was able to predict the change in behavior which occurred with increased confining pressure, provides corroboration of his hypothetical mechanism.

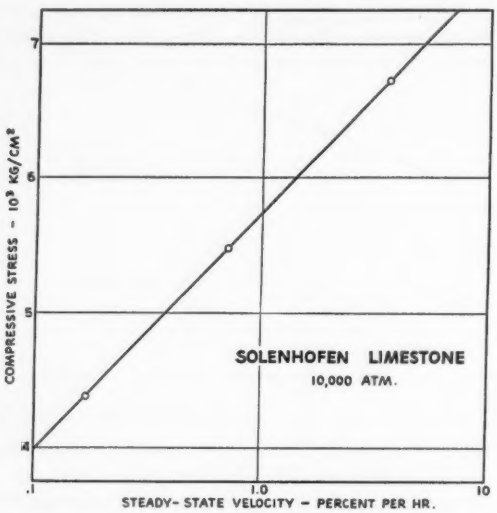

FIGURE 9.—*Creep rate of Solenhofen as a function of stress*
From creep tests under 10,000 atmospheres confining pressure

Another type of creep experiment showed this relation between stress and velocity of steady-state flow. Solenhofen limestone at room temperature, under a confining pressure of 10,000 atmospheres and with no solutions present, was tested in creep under high values of shear stress. The relation between the velocity of flow and the stress is shown in Figure 9. The data from these creep tests were published in an earlier paper (Griggs, 1936, p. 562). It is interesting that, although the conditions of these Solenhofen creep tests were about as widely different as could be chosen from those of the alabaster tests, the two nevertheless show a simliar relationship between velocity of flow and shear stress.

If it be found that this behavior is characteristic of pseudoviscous flow in general, then knowledge of three constants is sufficient to define this stage of flow for any given environment of confining pressure, temperature, and solutions: (1) the fundamental strength, (2) the slope

of the σ– log v curve and (3) the velocity at any one stress. From these relations it may be possible to develop equations of "plastodynamics" [2] similar in generality to those of hydrodynamics. It seems possible that this type of flow may prove to be most important in geological applications, and the development of a science of plastodynamics might well pave the way for the establishment of an exact science of rock deformation.

TABLE 1.—*Properties of alabaster in wet creep tests at different stresses*

(Gypsum in each experiment subject to constant load, in the presence of its own saturated aqueous solution, at 24° C., $\sigma_2=\sigma_3=0$)

Compressive stress ($\sigma_1-\sigma_2$ in kg/cm²) σ	Steady-state velocity (millionths per day) v_s	Deformation before fracture (per cent) S	Duration of test (to fracture) (days) t_f	Equivalent viscosity (poises) $\times 10^{16}$ η
300	2000	.89	2.45	.42
250	440	1.39	13.54	1.60
205	219*	2.3*	48*	2.64
181	100	3.0	133	5.11
165	77.0	3.6	285	6.04
150	66.5	>1.05	>110	6.36
125	24.5	3.8	308	14.4
103	7.0	>.71	>520	41.0 (?)

*Average of three duplicate tests.

Properties of same gypsum when dry:
Strength (short-time) = 520 kg/cm²
Deformation before fracture = .35% (primarily elastic)
Equivalent viscosity = > 5. $\times 10^{18}$ at 420 kg/cm²

DELAYED FRACTURE

One of the characteristics of the behavior of alabaster under these conditions of creep which has important applications to geology is the tendency of the specimen to fracture under a small stress that has been acting for a long period of time. This behavior was contrary to the writer's expectations, which were that the effect of solution and recrystallization under stress would be to dissolve material at the points of stress concentration and thus relieve the localized stresses inevitable in any crystalline aggregate. This would have the effect of rendering the stress distribution more homogeneous throughout the specimen, and consequently it was expected that under these conditions the material would be able to support a higher differential stress than in the dry state, instead of being much weaker.

[2] The term "plastodynamics" was suggested by C. W. Washburne.

As creep passes into the third stage and the creep rate increases, it appears that there is gradual loss of cohesion in the specimen, until finally complete separation occurs, producing sudden rupture very similar to normal breaking tests in the testing machine. It would seem that the increase in creep rate reflects the first stage in rupture, namely the beginning of localization of deformation along the surfaces which later develop into the surfaces of rupture, so we may say that incipient fracture occurs at the point of inflection where the creep rate begins to increase. Curiously enough, incipient fracture so defined is observed to occur at approximately the same amount of shortening, regardless of the stress or rate of creep. This is shown in Figure 6, where we see that in all the experiments, with the exception of those at 103 and 125 kg/cm^2, the creep rate begins to increase (curves upward from the straight line) between .45 and .55 per cent deformation. The significance of this "critical strain before rupture" is mentioned by Goranson (1940, p. 1030). It may be an important factor in the development of fracture cleavage simultaneously with flow cleavage. Thus it may be possible to get both flow cleavage and fracture cleavage in the same region by developing flow cleavage in those layers where the amount of deformation is below the critical value for fracture, and fracture cleavage where the deformation exceeds the critical value.

Another application of the principles illustrated in this flow may be made to rock fracture in earthquakes or in mine bursts. It might be expected, if deformation preceding rock fracture in these instances could be observed by geodetic or strain gage measurements, that the velocity of deformation would follow some such relationship as that shown in these creep experiments, namely an initial decrease in rate of deformation, followed by a period of more or less constant velocity and finally a period of increasing velocity. If such were the case, then it might be possible to develop criteria by which rock fracture could be predicted and some of the disastrous consequences avoided. Such prediction is possible in the case of laboratory creep experiments, and it does not seem beyond the realm of possibility that it may be developed for these other types of delayed fracture.

STRENGTH

These experiments serve as an interesting example of the difficulty of defining "strength" in laboratory tests. It is hard for one to overcome the idea that strength is a simple physical property which may be readily measured. The above observations show that the value of strength in these experiments depends on how it is defined. If we accept the definition that strength is exceeded when permanent set occurs, then we see that the limit of strength is set by the sensitivity of our measuring equipment

and by our definition of flow. For example, in the Solenhofen limestone creep test, it will take several years to determine whether the flow at 1400 kg/cm² is of the nature of permanent set or reversible elastic flow, and then we are not sure that a test of longer duration would not show a different behavior.

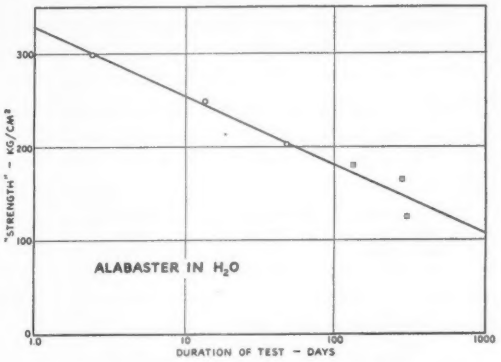

FIGURE 10.—*Strength of alabaster as a function of duration of test*

Data from experiments of Figure 6.

If we accept the definition that strength is that value of stress necessary to produce fracture of the material, then that strength is a function of the duration of the test. Figure 10 shows the relation between strength and duration of the test as determined from the experiments illustrated in Figure 6.

All the durations with which we have to deal are infinitesimal compared to geologic time. We certainly cannot extrapolate the curve of Figure 10 sufficiently to be of help in defining strength for geological purposes. Equation (1) suggests that a threshold strength exists (92 kg/cm²) below which the creep rate becomes zero and the specimen will hence support this stress indefinitely without fracturing. The difference in behavior of the test at 103 kg/cm² from those at higher stress may serve as an indication of what will happen when a stress slightly below the critical value is applied to the specimen. A hypothetical reconstruction is as follows: The alabaster will flow in the early stages very much as at higher stress, but the rate will gradually decrease to zero after a deformation of about .6 to .7 per cent, when nothing further will happen. If this should be so, then we have determined a value of strength which will have geological application.

This value of strength has previously been termed "fundamental strength" and defined as follows:

"The fundamental strength of a body is the differential pressure which that body is able to withstand under given conditions of confining pressure and temperature without rupturing or deforming continuously"

(Griggs, 1936, p. 564). It can be seen from this discussion what a time-consuming and difficult task it is to determine the value of the fundamental strength of a body under only one set of environmental conditions.

TABLE 2.—*Creep of alabaster at 1000 atmospheres confining pressure*

(Gypsum in each experiment subject to constant load, in the presence of its own saturated aqueous solution, at 24° C., $\sigma_2 = \sigma_3 = 1000$ kg/cm²)

Compressive stress (kg/cm²)	Steady-state velocity (millionths per day)	Deformation before fracture (per cent)	Duration of test (to fracture) (days)	Equivalent viscosity (poises) x 10^{16}
300	13,600	.470	.184	.064
250	2,650	1.5 ca	2.7	.271
200	390	....	>7.0	1.47

TESTS UNDER MODERATE CONFINING PRESSURE

With the aid of a recent grant from the Penrose Bequest, a new creep apparatus has been constructed which makes it possible to conduct tests of the type described above with the addition of one variable—confining pressure. The technique of experimentation is similar to that of the above tests except that the specimen is surrounded by a high-pressure cylinder and a pressure applied to the liquid which surrounds the specimen. In preliminary tests a liquid pressure of 1000 atmospheres has been used.

By analogy with previous high-pressure experiments which always showed an increased resistance to flow, the writer expected that this pressure on the liquid would cause slower creep and greater deformation before fracture. Just the reverse occurred, in agreement with Goranson's predictions. The increased solubility and mobility of the ions in solution prevailed over the other effect and resulted in a great increase in creep rate under similar conditions of differential stress. Further, the specimen showed a smaller amount of deformation before fracture under the same differential stresses, although for similar creep rate the latter experiments show a greater malleability than the former. The results are summarized in Table 2.

CONCLUSION

We have seen that high confining pressures may cause some rocks to flow with the production of textures resembling some of those found in nature. Pressure alone, however, does not reproduce all the conditions of rock flow in nature. It causes the strength to increase to values which seem too high to be attained by differential forces within the earth's crust. Moreover, it does not produce flow in quartz, as we have reason to believe occurs in nature. Because pressure alone has been insufficient, we must exercise great caution in applying the principles of deformation derived from high-pressure experiments to the deformation of rocks in the earth's crust. This is particularly true of the field of structural petrology, where one might conclude that mineral orientation produced under high pressure in the laboratory must be developed by processes similar to those which produce mineral orientation in nature.

Preliminary experiments indicate that temperature alone produces unimportant effects on deformation of rocks, unlike the metals, which change their properties markedly as the temperature is raised.

Creep tests show that the flow of limestone loaded to 1400 kg/cm^2 is so slow that the deformation would be negligible even if continued at this rate for millions of years.

When, however, conditions of testing are such that solution and recrystallization may occur, the observed characteristics of rock flow are entirely different. In marble, it is possible to produce deformation without the high development of mechanical twinning which seems to provide the mechanism of plastic deformation when the marble is deformed dry. In alabaster, where an inconsequential amount of deformation may be produced when the specimen is dry, the same specimen wet will flow at a relatively rapid rate.

It has been possible in the tests on alabaster to establish two principles of flow which may be of general application: (1) Fracture occurs after a certain amount of deformation, over a wide range of stress. (2) The rate of flow depends on the stress in a manner entirely different from ordinary viscosity. Thus, the logarithm of the velocity of deformation is proportional to the stress, so that as the stress is increased the rate of flow increases exponentially.

The goal of experimental investigation of rock deformation is the simultaneous attainment of all the conditions of environment which prevail in nature. The method of the experimenter is to investigate the effects of each variable separately and then their joint effect. Previous work has dealt with the effects of high confining pressure, high temperature, and varying velocities, together with some combinations of these

variables. The present paper adds experiments which illustrate the effects of solutions, combined with varying differential stresses, consequently varying velocities, and with confining pressure. The step remaining before we may say that the field has been covered in a reconnaissance way is the addition of high temperature to these systems.

With the aid of the recent grant from the Penrose Bequest of the Geological Society of America, apparatus has just been put into operation in which it is possible to deform rocks in the presence of solutions, under high pressures, with varying differential stresses, under high temperature, and for long periods of time. These conditions approximate metamorphic environments in nature. Preliminary experiments indicate important new effects, such as recrystallization flow of quartz aggregates. It is hoped that a survey of the new field thus opened up will provide some clues to the processes of dynamic metamorphism.

Although all such experiments may provide suggestive facts for geology, their value is limited without a theoretical interpretation of the mechanism of flow. Such an interpretation is suggested by Goranson in the following paper, and the correspondence between his theoretical relations and these experiments is encouraging.

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"FLOW" IN STRESSED SOLIDS: AN INTERPRETATION

BY ROY W. GORANSON

CONTENTS

	Page
Abstract	1023
Introduction	1023
Correlative phenomena	1023
Derivation of steady flow relations for a stressed solid in contact with a solvent	1025
Derivation of steady flow relations for a stressed metal	1027
Derivation of creep relations for relatively small stresses	1029
Initial and final deviations from steady state flow	1029
Flow relations under hydrostatic confining pressure	1030
Fundamental strength	1032
Works to which reference is made	1033

ABSTRACT

According to the hypothesis presented herein plastic flow in solids takes place by means of a change-of-phase transfer mechanism as solid→fluid→solid or solid→solution→recrystallization of solid. Expressions are derived from consideration of the thermodynamic potential relations for different physical conditions. The quantity that has been related to the viscosity of a fluid is thus seen to be a function of the "activation energy." If the hydrostatic confining pressure be sufficiently high then release of the energy of deformation stored up in the crystal lattice can take place only by a change-of-phase mechanism.

INTRODUCTION

The important problem in geophysics which concerns the mechanism of deformation and flow of rocks as a function of stress, temperature, heterogeneity, and hydrostatic confining pressure has interested the writer for many years. In fact, it was proposed some years ago to initiate a program of experimental work at the Geophysical Laboratory to test some of the hypotheses and ideas formulated as a result of thermodynamic reasoning. In the meantime David Griggs expressed his interest in carrying out experimental work along the same lines. However, enough experimental data are now available, largely through the outstanding work of Griggs, to indicate the validity of these hypotheses and, at his request, some of this material is therefore presented herein.

CORRELATIVE PHENOMENA

The main thesis of this paper is an interpretation of the mechanism involved in the phenomenon variously called "creep," "plastic flow" and

"pseudo-viscous flow." From the mechanism proposed herein are derived various formulae which satisfactorily correlate and fuse phenomena hitherto discussed as somewhat unrelated in treatments on strength, stress, strain, and strain rate.

A knowledge of the mechanisms involved in deformation is essential if generalizations made from studies of laboratory data are to be applied correctly to deformation phenomena in general or used as tools in acquiring further information.

Empirical relations connecting stress, strain, and strain rate are the boundary conditions which any theory must satisfy, but attempts to deduce a generalized analytical functional relation between these variables must fail because, as will be shown later, they are not in general sufficient to define the state of the system.

Volume compressibility, like density or heat capacity, is structure insensitive. By that we mean the volume compressibility of a single crystal is the same as that of a polycrystalline aggregate of the same material. Shear strain, creep, and other related phenomena, on the other hand, will vary with crystallographic orientation and other structural factors. Any relations we deduce must therefore be structure sensitive and thus will vary among different specimens of a polycrystalline aggregate.

Under certain conditions the material may react like a "perfectly elastic body" whereas under other conditions it may react like a "viscous liquid." In general our material will be found to behave in some intermediate manner for which the names "elastico-viscous" and "firmo-viscous" have been suggested. Other phenomena which must also be correlated are: (1) Strain hardening (work hardening) where the yield stress increases with strain, this coefficient $\partial X_y/\partial \epsilon$ being a constant equal to Young's modulus for "elastic bodies" and zero for liquids. (2) The change of yield stress with strain rate; this coefficient $\partial X_y/\partial v$ is a constant related to the coefficient of viscosity for liquids. (3) The type of deformation may also vary with the speed of loading, and a "ductile" substance may, under certain conditions of rapid loading, behave as a "brittle" substance. (4) Again "aging," which is presumably a slow transition toward a more stable state, has been observed wherein the yield stress changes with time. A related phenomenon may be observed in glass which, although breaking readily along a fresh scratch, breaks with more difficulty after a lapse of time and eventually will no longer break cleanly along the scratch.

Considerable experimental data and empirical relations are available for metal aggregates so that, if we can correlate rock and metal deforma-

tion, this large fund of data becomes valuable in the study of rock deformation. As will be shown, they differ only in degree, this difference being due to the variation in the crystal lattice bonding forces.

DERIVATION OF STEADY FLOW RELATIONS FOR A STRESSED SOLID IN CONTACT WITH A SOLVENT

The approach made herein on this problem is through thermodynamics. Several investigators (Bridgman, 1916; Williamson, 1917; Boydell, 1926) have recognized the validity of this tool, but its usefulness has, however, not been fully realized. One of the reasons for this may be that valid thermodynamic expressions cannot be derived from the actual physical conditions. Boydell, although he sets up the problem of an unequally loaded solid, actually derives Poynting's expression which applies only to the case where hydrostatic pressure p_1 acts on the solid and p_2 on the liquid, $p_1 > p_2$. Although, coincidentally, Poynting's expression has the same form as (1A) of this paper this fact still does not justify his resulting argument.

The type of flow or creep phenomena observed in metals is not, in general, apparent in rocks because the cohesive bonding forces are much higher for rock-forming minerals than in general for metals. This type of flow may, however, be simulated very closely experimentally if the rock be surrounded by a liquid in which it is somewhat soluble. Thus, instead of beginning with the simpler system, the relations will be derived first for the latter system because the physical mechanism is here more readily perceived.

Assume an initially "ideal solid," namely, one in which the thermodynamic potential is the same for all the faces and equal to the potential of the solid in solution (solute)—*i. e.*, the solid is in equilibrium with the saturated solution at a temperature T and hydrostatic pressure P. If now the solid is loaded by a longitudinal compressive stress X the thermodynamic potentials at the stressed and at the free faces of the solid are no longer equal and the system is no longer in equilibrium so long as the stress exists.

In order to derive our equations the system is therefore first subdivided into hypothetical isolated parts, namely the regions comprising the stressed surface, the free face, and the solution bulk. We also assume any coexistence of phases necessary for our derivations. The physical interpretation follows readily from this procedure, and we furthermore avoid any confusion or error that might result from a compromise between the thermodynamic and the physical pictures.

The following formulae were derived by ordinary thermodynamic methods applied to stressed systems (Goranson, 1930, 1937) and have been somewhat simplified for purposes of clarity. We have

$$\left(\frac{\mathrm{d} \ln a_s}{\mathrm{d} X}\right)_{S.F.} = \frac{M}{\rho RT} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (1A)$$

and

$$\left(\frac{\mathrm{d} \ln a_s}{\mathrm{d} X}\right)_{F.F.} = \frac{MX}{\rho ERT} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (1B)$$

where equation (1A) refers to the stressed face (subscript *S.F.*) and (1B) to the free face (subscript *F.F.*). a_s denotes the activity of the solute which, for an ideal solution, is equal to the mole fraction concentration of the solute in the solvent. X denotes the compressive stress, M the effective mole weight, ρ the density of the solid, R the gas constant ($R = 83.156$ *bar cm*), and T the absolute temperature. E is Young's modulus of elasticity in compression—*i. e.*, $Ee_1 \doteq S_1$ where S_1 is the stress (negative for pressure) and e_1 the extension per unit length in the direction of the stress.

On integrating (1A) we have at the stressed face

$$X - X_0 = \frac{\rho RT}{M} \ln (a_s/a_{s_0})_{S.F.} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (2)$$

In the system gypsum and water at 24° C. the rate of solubility increase at the stressed face under compressive load is threefold per kilobar. The rate of solubility increase at the free face will be $3X/E$ which, for $E = 10^6$ *bars* and $X = 1000$ *bars*, is only one-thousandth of the rate of the stressed face.

Thus at the stressed face solubility is increased by a compressive stress and lowered by a tensile stress. At the free face solubility is increased by either a compressive or tensile stress but by a much smaller factor at moderate stress. (However, *see* equation (9).)

Now, whereas under compressive load the solute concentration tends to increase at both stressed and free face, diffusion of solute away from these faces would leave the bulk of the solution supersaturated. Consequently deposition of solute should occur at unstressed places. If, however, the solution can supersaturate by an amount in excess of the relatively small increased solubility at the free faces, then crystallization will take place on these free surfaces which here act as nuclei for deposition. Our mechanism can thus be considered as a diffusion process along crystal grain boundaries from loci of high to low stress wherein the solution acts as a transfer medium. This diffusion rate will approach a steady state when a dynamic equilibrium between rate of solution and

rate of deposition has been established and this rate will be governed by the mobility of the solute in the solution, the path length, and the concentration head between the stressed and free faces.

If by varying the stress we change only the concentration head then the change in the steady state creep rate, v_s, should be directly proportional to the change in activity or

$$\frac{\mathrm{d} \ln a_s}{\mathrm{d} X} = B \left(\frac{\mathrm{d} \ln v_s}{\mathrm{d} X} \right) \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (3)$$

where B is a constant. From (2) and (3) we have

$$X - X_o = K \ln (v_s/v_o) \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (4)$$

where K, X_o, and v_o are physical constants. This is the empirical relation Griggs (1940) obtains for the steady creep rate of stressed gypsum in contact with water.

On increasing the compressive load the solubility at the stressed surfaces eventually becomes larger than the amount by which the solution can supersaturate. Under these conditions the solution will deposit the excess solute in any available unstressed space. In consequence the texture of the polycrystalline substance may become so loosened that it crumbles under the load.

DERIVATION OF STEADY FLOW RELATIONS FOR A STRESSED METAL

Because the remainder of the argument is closely paralleled by that for "plastic" flow in metals the relations for these substances are inserted herein. It should be emphasized, however, that these relations are not limited to metals but apply also to other solids as, for example, limestone (Griggs, 1940, Fig. 9).

We shall, as before, divide our system into hypothetical isolated portions and derive expressions for the stressed and free face. We assume also the coexistence of two phases—the bound atoms of the solid and the free migrating atoms with energies exceeding those necessary to break the cohesive bonds of the crystal lattice. At the melting point this will be equivalent to the heat of melting.

The expressions analogous to (1A) and (2A) connecting compressive stress X and the melting temperature T_m in degrees absolute are

$$\left(\frac{\mathrm{d} \ln T_m}{\mathrm{d} X} \right)_{S.F.} = -\frac{1}{\rho h} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (5A)$$

and

$$\left(\frac{\mathrm{d} \ln T_m}{\mathrm{d} X} \right)_{F.F.} = -\frac{X}{\rho E h} \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots (5B)$$

where h denotes the heat of melting, and the other quantities are as before. In the following discussion this quantity h is used as an activation energy of the process.

The melting point at the stressed face is thus depressed for compression and raised for tension. The melting point at the free face is depressed for both compression and tension. The relative lowering of the two faces is the same as that found between (1A) and (1B).

The number of atoms with energies in excess of that needed to break the lattice bonds—*i. e.*, the number of free migrating atoms—increases with increase in temperature, and at the melting point this fraction of the total number becomes unity. The factor

$$\theta = (T/T_m), \quad (6)$$

where T denotes the experimental temperature and T_m the melting temperature, has been used in correlating temperature versus creep data for different metals. The behavior of a low melting solid is thus considered as equivalent in behavior to a high melting point substance at an experimental temperature correspondingly higher for the same value of θ, other factors remaining the same. In other words, lowering the melting temperature has the same effect as raising the test temperature. This is found to be a sufficiently good approximation for metals but should be carried out in terms of energy if different types of lattice structures are intercompared. When this is done one obtains a more complicated expression which is omitted here because it is not essential for our present purposes but will be given in a later paper in connection with some other work.

On combining equations (5A) and (6) and integrating we obtain

$$X - X_o = \rho h \ln (\theta/\theta_o) \quad (7)$$

For a sufficiently high stress—*i. e.*, a relatively large creep rate—we may in the same manner as before write

$$X - X_o = K' \ln (v_s/v_o) \quad (8)$$

where K' is a physical constant and v_s the steady state or minimum creep rate. This is the expression that has been used for about 30 years to express the empirical relations between yield stress X and minimum creep rate v_s for metal aggregates.

It should be noted that the integral expression of (5B), and similarly of (1B), yields the equation

$$\frac{X^2 - X_o^2}{E} = K'' \ln (v_s/v_o) \quad (9)$$

which may become large for large X. This is the plastic flow relation for tension.

DERIVATION OF CREEP RELATIONS FOR RELATIVELY SMALL STRESSES

Expressions (4) and (7) were derived for relatively large creep rates and thus for compressive loads large enough to iron out the initial high inequalities over a short period of time. Let us assume now that we are operating with either very small stresses or with a substance that does not "migrate" readily. For these conditions the number of migrating units will be relatively few and their paths much shorter. Thus, instead of being able to picture a statistical streaming action we are slowed down to a discontinuous hop-skip process and can no longer set up a steady mean. The crystallization or freezing process must be explicitly considered here, and for these cases we have, for migration at the stressed surface

$$K\,(\theta/\theta_o) = K\,e^{\left(\frac{X - X_o}{h\,\rho}\right)}$$

and for fixation at the stressed surface

$$K\,(\theta_o/\theta) = K\,e^{-\left(\frac{X - X_o}{h\,\rho}\right)}$$

The creep rate will be given by the difference between these two quantities or

$$(v/v_o) = K\left[e^{\left(\frac{X - X_o}{h\,\rho}\right)} - e^{-\left(\frac{X - X_o}{h\,\rho}\right)}\right]$$

$$= K \sinh\left[\frac{X - X_o}{\rho\,h}\right]$$

The same reasoning, and therefore a similar sinh expression, also replaces (4) for very small creep rates.

An expression of this type was obtained empirically (Mussmann, 1938) by combining analytically the empirical logarithmic relation for high creep rates (7) with the linear relation between stress and creep rate observed at very low creep rates.

INITIAL AND FINAL DEVIATIONS FROM STEADY STATE FLOW

Thus far we have considered only the relation between steady creep rate and stress. When compressive load is first applied the irregularities in the structure of the polycrystalline material set up an initial localization of stress at the raised points and thus, for large enough loads, an initially high creep rate. This rate gradually diminishes toward a steady state as the original inhomogeneities of texture become ironed out

and the stress redistributes itself over larger and larger surface areas. The effective stress, for constant load, thus decreases with time. Finally the pore spaces in the texture become filled, and the grains become more or less reoriented into their most stable crystallographic configuration for an axial load distribution. If now a series of such creep-time tests is made for varying compressive loads on initially identical specimens this last state should be reached at about the same longitudinal contraction ϵ_m, assuming none of the material has precipitated elsewhere. At this point the specimen has stabilized itself and from this point on should therefore behave more or less as a unit—*i. e.*, approximate to the condition for a single crystal. Solution (or migration) then begins to work along crystallographic surfaces and will be most effective on surfaces across which the cohesive bonds are smallest—*i. e.*, have the largest free energy, provided they are in a general 45-degree trend to the stress axis. If the load is large enough the specimen will then deform by a gliding action along 45-degree planes, and the creep rate, measured by the rate of contraction, will appear to accelerate until the specimen fails (Griggs, 1940, Fig. 6).

Plastic flow may also occur in single crystals, but the effect in general is a gliding along cleavage, twinning, or parting planes from combined "melting" and snapping of bonds; twinning is a definite stabilization phenomenon. Localized high stress regions may be set up in crystals, as was mentioned earlier, and thus an initially single crystal may, under a compressive load, end up as a mosaic of reoriented crystalline grains.

FLOW RELATIONS UNDER HYDROSTATIC CONFINING PRESSURE

It was also mentioned earlier that the rate of flow by the mechanism of solution and recrystallization will be a function of the concentration head, mobility of the solute in the solution, and the path length. If we vary not only the longitudinal stress but also the hydrostatic confining pressure we shall also vary the solubility relations, the concentration head between the stressed and unstressed faces, and the mobility of the solute. The most important single factor is the mobility of the solute. Now the validity of equation (4) depends on the applicability of (3), but in expressing the latter we assumed only a change in the concentration head due to a change in the unidirectional load. This we can no longer assume if we also vary the hydrostatic pressure. The solubility of alabaster in water increases with pressure up to some pressure value in excess of 1000 atmospheres. What is more important to us is the fact that the mobility also increases under pressure to some pressure value also in excess of 1000 atmospheres.

In an attempt to predict the effect of a confining pressure of 1000 atmospheres on the rate of flow of a compressively stressed alabaster specimen in contact with water the ratio

$$\frac{\left[\frac{\mathrm{d} \ln v_s}{\mathrm{d} X}\right]_{1000 \text{ atm.}}}{\left[\frac{\mathrm{d} \ln v_s}{\mathrm{d} X}\right]_{1 \text{ atm.}}}$$

was computed to be about 1.5 from the expected increase of about 50 per cent in the activity and mobility of the solute. Griggs' experimentally determined factor is about 1.6. The closeness of this agreement is coincidental because lack of proper data will permit in general only an order of magnitude check.

It might be expected that the effect of increased confining pressure, by forcing liquid into and through the pores and cracks of the solid, will increase the effective surface area and so result in a higher initial creep rate until these open spaces become filled with deposited material.

The energy at the external crystal faces cannot be denoted by the energy of the interior because of loss of symmetry at these external faces; the difference is known as surface tension. For ionic crystals this difference will be greatest along edges and at corners and least at the centers of faces; this is the reason for the initial skeleton growth of such crystals. With a high symmetry type of structure this difference might become greatest at re-entrants. Similarly for the systems we have been considering solution or melting will be most rapid at the stressed surfaces of highest free energy, and growth will be most rapid on the free faces of lowest free energy. These directions are, in general, indicated by the crystalline form and cleavage. For example mica and related minerals, which form planar structures held together by feebler bonds, tend to grow with their cleavage planes perpendicular to the axis of greatest compressive thrust; asbestos and related fibrous minerals, which form chain structures bonded together by feeble forces, tend to elongate in the plane perpendicular to the thrust.

This type of stress structure is very striking when more than one kind of bond is present in the crystal structure of the component minerals, as for example in mica schists. Stress structure is apparent also, though to a lesser extent, even where the crystal bonding is much more symmetrical in space, as in the observed reorientation of quartz grains.

A theoretical discussion which attempts to predict the effect of combined compressive stress with high hydrostatic confining pressure acting simultaneously on a solid has been presented by the writer (Goranson,

1940). According to this hypothesis permanent deformation of a solid will take place only by the change of phase mechanism postulated above, provided the confining pressure be sufficiently high. The hypothesis may not be correct in all its details, but it does seem to be strikingly corroborated by the work of Griggs and by some recent work of the writer.

FUNDAMENTAL STRENGTH

The term "strength of the material" has been employed with so much looseness that it has very little significance unless its usage is defined wherever employed. The "yield stress" is supposed to denote the differential stress below which the material will always recover its original shape and dimensions when the stress is released. The yield stress is ordinarily determined, however, by a short time test which yields values that are too high because the true yield stress can be obtained only by extrapolation to infinite time, or, in practice, by extrapolation of stress versus creep rate curves to zero creep rate. Griggs uses the term "fundamental strength" as synonymous with this true yield stress. At absolute zero temperature the fundamental strength will be determined by the potential or elastic strength—*i. e.*, the stress for rupture. At other temperatures the solid might deform by the phase-change transfer mechanism. Factors favoring the operation of this mechanism are nonuniformity of stress distribution in the material, nonisotropy in crystalline bonding, and high temperature.

Given infinite time, there is a statistical probability of a lattice point acquiring sufficient thermal energy to break free from its bonds and migrate, but the firmer the lattice bonds and the lower the temperature the greater must be this activation energy. Experimentally a stress will be found below which no measurable flow will occur in any finite time—*i. e.*, the energy barrier will in practice be too high. This stress value will then denote the fundamental strength of the material for the temperature and other conditions of the test.

The fundamental strength is therefore a function of the elastic potential and thermodynamic potential relations.

Rock flow of importance to geology may take place under conditions which are so widely different from those obtainable in the laboratory as to call in question any direct comparison or application. In other words, it has not yet been possible to duplicate these processes by any single laboratory model. For example, the activation energy of quartz will be greater at room temperature by an order of magnitude (about 10 times) than at depths of hundreds of kilometers, and thus flow phenomena will be very different for these two conditions. The time available for

geological processes is also so much greater than that in laboratory tests as to render extrapolation of our meager data very uncertain.

On the other hand, laboratory data supply the building blocks upon which it is possible to formulate and check generalizations in terms of physical hypotheses. Only after all the variables concerned have been taken into consideration can the results be applied or extrapolated with any assurance.

Griggs and the writer may perhaps differ in certain still speculative aspects of dynamic geology but they are working together in perfect agreement that advances will be made by attempting to formulate into a generalized hypothesis the mechanism of flow and rupture co-ordinated by experimental data.

It is a pleasure to acknowledge the benefit the writer has received throughout his long association with Mr. Griggs from discussions which have in large measure contributed to the formulation of the hypothesis here advanced as to the mechanism of flow.

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EQUILIBRIUM BETWEEN VAPOR AND LIQUID PHASES IN THE SYSTEM CO_2—H_2O—K_2O—SiO_2

BY GEORGE W. MOREY AND MICHAEL FLEISCHER

CONTENTS

	Page
Abstract	1035
Introduction	1035
Acknowledgment	1037
Apparatus and methods	1037
Temperature measurement and control	1040
Materials	1042
Experimental procedure	1043
Calculation of results	1046
Attainment of equilibrium	1048
Rate of reaction of quartz with carbonate solutions	1048
Discussion of results	1051
Summary	1057
Works to which reference is made	1057

ILLUSTRATIONS

Figure	Page
1. Apparatus	1038
2. Preparation of mixtures	1042
3. Rate of solution of quartz	1050
4. Compositions of coexisting phases	1052
5. Distribution of carbon dioxide	1053
6. Composition of vapor	1054
7. Distribution of water	1055

ABSTRACT

Experimental results in systems which contain both nonvolatile and volatile components, and in which the vapor phase is complex, are needed to increase our knowledge of processes considered to be of importance in ore deposition. The major difficulty in such a study is one of technique. An apparatus has been constructed which makes possible the separation and independent analysis of liquid and vapor phases.

The new apparatus has been tested in a study of the system H_2O—CO_2—K_2O—SiO_2, the results of which are presented in tables and curves.

INTRODUCTION

Chemical reactions that take place by the agency of hot aqueous solutions and hot compressed gases, known as hydrothermal and pneumatolytic processes, respectively, are of primary importance in the formation

of many ore deposits. Concerning the mechanism of these processes, and especially concerning the mechanism of transport of material and both the relative and absolute importance of transport in liquid and vapor phases, there is no unanimity of opinion. Excellent presentations of differing views have recently been made by Fenner (1933) and by Ross (1933). Although the limited available physicochemical data have been used freely in the formulation of these opinions, nearly all these data refer to properties of pure substances, and their application to the multicomponent systems involved in geological processes is of uncertain validity.

Such uncertain extrapolations can be avoided only when direct experimental results become available. The need for research on the physical chemistry of systems containing volatile components at high temperatures and pressures has long been recognized, but the difficulties resulting from the extreme experimental conditions have delayed the attack on such systems. Recently, however, the study of the fusion—solubility relations in parts of the system H_2O—Na_2O—K_2O—Al_2O_3—SiO_2, begun many years ago by Morey and Fenner (1917), has been resumed in the Geophysical Laboratory by Morey and Ingerson (1937a, 1938) and by Morey and Burlew (1938). It is hoped that these investigations will be of service in elucidating problems arising from the pegmatitic and late hydrothermal stages of igneous activity.

Field studies of ore deposits have shown that many ore-bearing solutions contain more than one volatile constituent, and numerous questions have been raised thereby that cannot be answered satisfactorily at present because pertinent physicochemical data on such systems are lacking. For instance, the orderly sequence of deposition of certain minerals in many ore deposits has been explained (1) By differential precipitation from an originally complex solution, or (2) by precipitation from a solution relatively simple in composition at any given time but constantly changing in composition at the magmatic source. Plainly, knowledge of the physical chemistry of such solutions must be widely extended before it will be possible to pass upon the validity or competency of the suggested mechanisms.

It is evident that the deposition of minerals from solutions, whether those solutions be liquid or gaseous, is determined by temperature, pressure, and concentration factors. Relative volatilities have frequently been invoked, sometimes without the realization that the partial pressures of the constituents of a complex vapor may be different even in order of magnitude from the vapor pressures of the pure volatiles. The same is true of relative solubilities. Accurate prediction of the behavior of such solutions can be made only after obtaining a comprehensive knowledge not only of the solid–liquid equilibria (the solubility relations)

but also of the effects of temperature and pressure on the distribution of volatiles between the liquid and vapor phases in multicomponent systems.

The investigation here reported is the first attempt to obtain the latter type of data experimentally. The problems to be answered may be stated as follows: (1) Given a mixture of known temperature, pressure, and composition, how will the volatiles be distributed between the liquid and vapor phases? (2) Given a solution of known temperature, pressure, and concentration, how is the distribution of the volatiles between the liquid and vapor phases affected by reduction in pressure at constant temperature?

The solutions involved in ore deposition are extremely complex in composition, probably containing more or less of all the common rock-forming elements as well as a number of volatile constituents. It is not feasible at present to undertake the experimental study of such complex solutions. The number of components had to be reduced to simplify the problem to a point where it could be handled. Accordingly, we have restricted our work to the four-component system K_2O—SiO_2—CO_2—H_2O. Water and carbon dioxide are two of the most important volatile constituents of magmas. Furthermore, the other possible choices such as the halogen or sulfur compounds would have greatly increased the experimental difficulties due to corrosion. K_2O and SiO_2 were chosen as the other components because the high solubility of the potassium silicates in water allows working at relatively low pressures and because the relations in the bounding ternary system were known from the work of Morey and Fenner (1917).

ACKNOWLEDGMENT

The problem and its method of attack were formulated at conferences between Arthur L. Day, G. W. Morey, C. H. Behre, and T. S. Lovering in 1935. It is a pleasure to acknowledge financial aid from two grants which were made to T. S. Lovering from the Penrose Bequest of the Geological Society of America to carry out the investigation devised by this group. The work, which was carried on in the Geophysical Laboratory by Dr. Fleischer and Dr. Morey, was financed by these grants for 2 years and for a third year by the Carnegie Institution of Washington.

APPARATUS AND METHODS

The study of the distribution of two volatiles between a liquid phase and a vapor phase requires the determination of the total amount and composition of each phase after equilibrium has been reached, or, what is equivalent and experimentally simpler, determination of the total

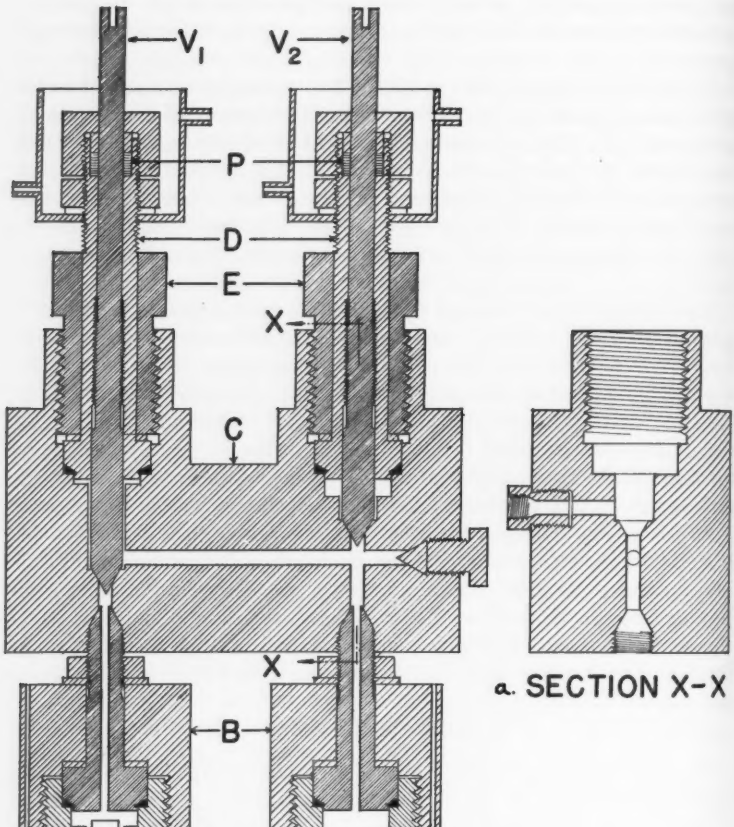

FIGURE 1.—*Apparatus used*

The two bombs A_1 and A_2 are connected through a valve-block C, provided with valves V_1 and V_2, which make possible the separation and analysis of vapor and liquid phases.

amount and composition of one of the phases and knowledge of the amount and composition of the original mixture. Sampling the phases is permissible only if the proportion of the total amount taken in the sample is known. Since it is experimentally impracticable to sample the liquid phase, it is necessary to use some method for separating the phases and subsequently analyzing one or both.

In the early work on the system, the hydrothermal quenching method developed by Morey and Fenner (1917) and used successfully by them in studying the system K_2SiO_3—SiO_2—H_2O was tried. In this method, the phases are separated by chilling the bomb in such a way that the liquid phase is "frozen" to a glass, while the water in the vapor phase condenses on the walls of the bomb away from the "frozen" liquid phase. The latter can then be weighed and analyzed. Because this method is relatively simple and direct, considerable time was spent on it. Carrying out the quenching operation properly is something of an art, but with experience this method could be used with some success in the silica-rich portion of the system. However, when it was found that mixtures low in silica held too much carbon dioxide in the liquid phase to be quenched successfully, the method was abandoned.

The method finally adopted consists of separating the phases by means of a valve and then analyzing the vapor phase. In practice, a quantitative separation is not possible, and all the liquid phase plus a small part of the vapor phase is separated from the greater part of the vapor phase. As the volumes occupied by the vapor phase on each side of the separating valve can be determined, this procedure amounts to removing and analyzing a known proportion of the vapor phase. The total amount and composition of the vapor phase are then obtained by multiplication of the analytical results by an accurately determinable factor.

The apparatus consists essentially of two bombs whose inner chambers are connected through an opening within a valve block. The design is probably apparent from Figure 1, which shows a cross section through the center of the apparatus. The bombs A_1 and A_2 are of the type recently described by Morey and Ingerson (1937b), differing only in the sizes of the inner chambers and in that the plungers are hollow. The pressure seals are of the imprisoned copper washer type used for many years in this Laboratory. The annealed copper washers, shown in black in Figure 1, are imprisoned between the plunger, which is cut with a 60-degree taper, and the seat. The bombs are connected to the valve block C by 60-degree cone fittings on the plungers.

The valve assemblies are identical except for the length of the valve stems V_1 and V_2, both of which are shown in the closed position in Figure 1. The stems are threaded into the plungers D, D, which are tight-

ened by means of the threaded plugs *E*, *E*. The pressure seals are of the same type as those in the bombs. The valve stem packing *P*, *P* consists of alternate layers of oiled leather and synthetic rubber (Neoprene), which are held tight by the packing nuts shown in Figure 1. The packings are protected from the effects of high temperatures by circulating water through the outer jackets shown in the same figure. Attempts were made to obviate the necessity of water-cooling by using metal or asbestos or graphite-asbestos packing that would be resistant to high temperatures, but all the materials tried failed to hold pressure. The valve stems seat in both the open and closed positions, so that the valve stem packing is exposed to high pressure only while the valve is being opened or closed.

Valve V_1 shuts off the liquid phase in the crucible in bomb A_1 from the greater part of the vapor phase, in bomb A_2. Valve V_2 is used to withdraw the vapor phase for analysis. Figure 1, a, showing the section marked *X*-*X* in Figure 1, gives the details of valve V_2 and the vapor outlet. A steel tube 9 inches long (not shown in Figure 1, a) connects the vapor outlet to the analytical train.

All the threads shown are lubricated with a graphite-cylinder oil suspension. Graphite does not bake into stainless steel but forms a lubricating film between the steel parts after the oil has burned away. The entire apparatus is constructed of stainless steel (Carpenter No. 8), excepting the screw caps *B*, *B* of the bombs and the threaded plugs *E*, *E*, both made of tool steel (Halcomb Ketos) to avoid seizing, which occurs frequently if two stainless steel parts are screwed together under considerable tension. In the apparatus, the vapor phase comes in contact only with stainless steel, which is not appreciably corroded by steam up to approximately 500°, but above 500° scaling becomes noticeable.

TEMPERATURE MEASUREMENT AND CONTROL

The temperature that determines the equilibrium in the system is that of the liquid phase, and the equilibrium conditions are not affected if the vapor phase is not at the same temperature as the liquid phase, provided that no part of the system is at a temperature at which water can condense. The condensation temperature varies with pressure, but since water cannot be liquefied at temperatures above its critical temperature, 374°, this is the minimum safe temperature. The requirements to be met by the heating unit are, then, that the liquid phase in bomb A_1 can be held at a desired constant temperature and that the vapor phase, in the valve block opening and in bomb A_2, can be heated to a temperature above 374°. Furthermore, in designing the heating units, allowance had to be made for the very large heat loss up the valve

stems, a heat loss made inevitable by the water-cooling of the packings.

The heating arrangement finally adopted consists of two independent units. One of these is a furnace containing two heating elements, one for each bomb. Each of these is an Alundum cylinder, of size to fit snugly over a bomb, wound with Nichrome wire, of resistance 13.5 ohms. These are connected in series and well insulated with magnesia. The furnace is connected to the laboratory 110 volt regulated A.C. supply, with a Variac (a variable transformer built by the General Radio Company) and a Leeds and Northrup temperature controller in the circuit. The controller is actuated by a double platinrhodium vs. gold—palladium thermoelement placed in the furnace between the two coil heaters. The controller operates a make and break switch, which is connected in parallel to a 7.5 ohm resistance in the heating circuit.

The second unit of the heating arrangement is a pad heater, for independently supplying heat to the valve block. This was made of two Transite pieces, 15 by 8 by 1 cm., which fit closely against the two long sides of the valve block. Each pad has imbedded in it a Nichrome coil, and the two are connected by a short coil, originally of Nichrome wire, but later changed to silver, which is less brittle and less likely to break. The total resistance of the unit is 31 ohms. The pad heater is connected to the laboratory 110 volt regulated A.C. supply through a Variac. The furnace was covered above the pad heater with an asbestos cover, which reduced the heat loss by radiation.

As indicated above, a constant current was passed through the pad heater, only the current through the furnace coils being regulated. It was found desirable to put about 65 per cent of the total energy input into the pad heater. With the bombs at 500°, this kept the interior of the valve block at 400 ± 5°, as determined by a series of special runs with thermoelements inserted in place of the valve stems.

The very large change in e.m.f. of the regulating thermoelement (approximately 120 microvolts per degree) and the lagging effect of the large mass of steel in the furnace serve to keep the temperature inside the bombs quite constant. Overnight, the temperature was usually constant to ±0.4°, and during the last hours of a run the temperature was held constant to ±0.2° by adjusting the regulation if necessary.

Temperatures were measured with Pt vs. Pt—10% Rh thermocouples placed at the bottoms of wells 5 cm. deep drilled into the screw caps *B*, *B* of the bombs. The wells are shown in Figure 1 but are 30 degrees from their actual positions, the bombs being assembled so that the wells are directly below grooves cut in the pad heater. A special series of runs, made with thermocouples inside the bombs in place of the valve stems and with the platinum crucible in bomb A_1 half full of NaCl, showed

that the temperatures measured in the well differed from the actual temperatures of the crucible's contents by −0.1 to ±0.2°. The electromotive force from the thermocouples was converted to temperature by means of the table given by Adams (1914) and a deviation curve obtained

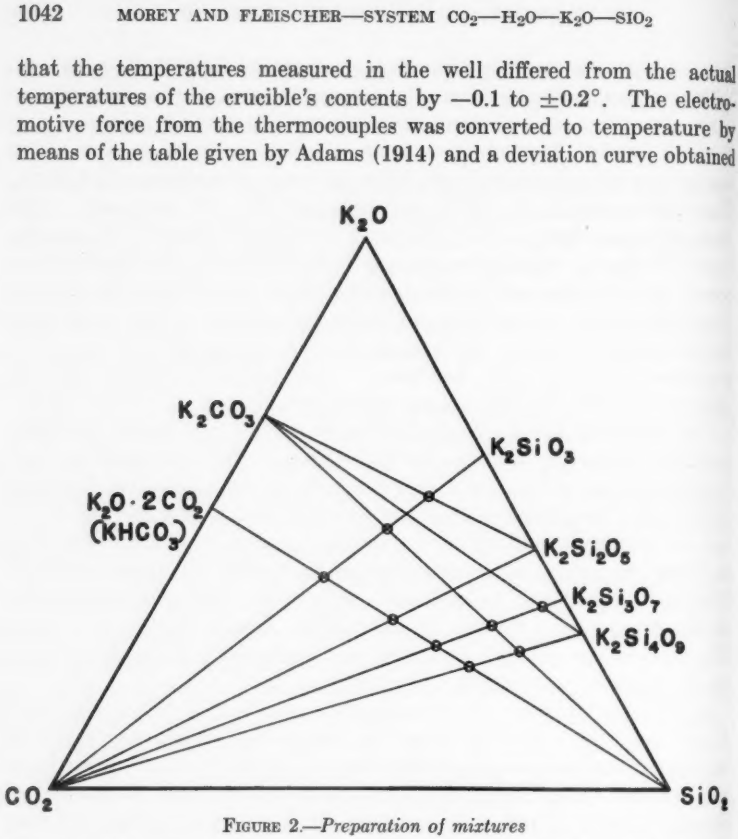

FIGURE 2.—*Preparation of mixtures*

Compositions of a given ratio of $K_2O:SiO_2$ could be prepared from $KHCO_3$ or K_2CO_3, and from SiO_2 or $K_2Si_4O_9$, thus introducing differing proportions of CO_2.

by frequent calibration at the melting or freezing points of ice, tin, zinc, antimony, and aluminum. One pair of thermocouples was used for nearly 2 years and showed only small changes, indicating that they were not contaminated by contact with steel at 500°. Electromotive forces were measured on a Wolff potentiometer, fully shielded and equipped with a White eliminating switch.

MATERIALS

The materials used were $KHCO_3$, K_2CO_3, SiO_2, $K_2Si_2O_5$, $K_2Si_4O_9$, and water. Baker's Analyzed $KHCO_3$ was found to be pure and was merely crushed and stored in a desiccator. K_2CO_3 was prepared by heating

$KHCO_3$ at 220° for 24 hours. Conversion was quantitative, as proved by determination of the loss in weight and by analyses of the product. This method is preferable to direct crystallization of the carbonate, which gives $K_2CO_3 \cdot 1\frac{1}{2}H_2O$, the hydrate stable up to about 235°. Dehydration of this compound is not so smooth as the decomposition of the bicarbonate. The SiO_2 used was quartz from Baker and Adamson. An analysis of this material by J. W. Greig of this Laboratory showed that a residue of 0.04 per cent was left on evaporation with HF and H_2SO_4. It was sieved, and the portion passing an 80 mesh sieve was used. The water used was distilled water that had been boiled 15 minutes while CO_2-free air was bubbled through it. The potassium silicates were synthesized from $KHCO_3$ and SiO_2. They were fused repeatedly until homogeneous glasses were obtained, then crystallized at appropriate temperatures, and stored in a desiccator over KOH. Their compositions were known from the synthesis and checked by analyzing them for K_2O by evaporating with HF and weighing as K_2SiF_6.

By the use of varied original ingredients a given ratio of K_2O to SiO_2 could be obtained with differing amounts of CO_2, and all the CO_2 was introduced as carbonate or bicarbonate. Thus (Fig. 2) a mixture of metasilicate ratio was obtained from $KHCO_3$ and SiO_2; K_2CO_3 and SiO_2; and K_2CO_3 and $K_2O \cdot 2SiO_2$. In future experiments it is planned to increase the proportion of CO_2 by introducing it as gas under pressure, for which the design of the apparatus is well adapted.

EXPERIMENTAL PROCEDURE

The solids used in making up the charges were weighed into the platinum crucible. Water was then added from a weight burette, the solids in the crucible being wetted down and the rest of the water being put into bomb A_2. The bombs were then assembled and tightened hard with a 21-inch wrench. The valve block and plunger assembly was then completed, all threads having been lubricated in advance with a graphite-cylinder oil suspension. The bombs A_1 and A_2 were then tightened again. This was a necessary precaution. It was found, after considerable trouble had been experienced with very slow leaks, that the copper gasket seals within the two bombs could best be made to hold by tightening them hard, allowing the gasket 20 to 30 minutes to flow somewhat, then retightening. After this procedure was adopted, no trouble was had with leaks at these gaskets.

After the plungers of the bombs were screwed into the valve block and were tightened up slightly on the cone seats, the vapor outlet tube was connected, the measuring thermocouples put into the wells, the pad heater slipped on around the valve block, and the bombs lowered carefully

into the furnace. Valves V_1 and V_2 were opened, the bombs evacuated through the vapor outlet by means of a water pump to a pressure of 50 mm., and valve V_2 was closed. This partial evacuation was made to reduce the amount of CO_2 formed in the apparatus by reaction of graphite or carbonized oil with the oxygen of the air within the apparatus. Blanks run without evacuating the apparatus give 0.009 g. CO_2, in good agreement with the value calculated on the assumption that all the oxygen within the bomb was converted to CO_2. Blanks run after evacuating to 50 mm. gave 0.0015 ± 0.0005 g. CO_2, somewhat more than the calculated amount. The analyses were corrected by subtracting 0.0015 g. from the analytical results for CO_2. During the evacuation, water is lost. The amount was determined by evacuating through a weighed Anhydrone tube, and this amount was subtracted from the original weight of water added to the bombs. It was usually less than 2 milligrams.

After the bombs were evacuated and valve V_2 closed, a slow stream of water was started through the jackets around the valve packing, and the apparatus was heated, very slowly at first. Usually, the rate of heating was such that 2 hours was required to raise the temperature to 150°; the current was then increased, and the equilibrium temperature, 500°, was reached in an additional 2 to 3 hours. The slow initial rate of heating prevented too rapid evolution of CO_2 and the consequent spattering. When equilibrium was approached from the low temperature side, the usual procedure, the regulator was set and started when the temperature had nearly reached 500°, and the run was held at 500° overnight. The following morning, temperature readings were made every 10 minutes for 2 to 4 hours before analyzing.

In order to approach equilibrium from higher temperatures, the Variac in the furnace line was set to give a slightly higher temperature. The next morning, the regulator was turned on, and the apparatus cooled very slowly to the equilibrium temperature, 500°. Rapid cooling or too great a temperature drop caused a leak. It was not safe usually to cool more than 30 degrees, although one successful run was made in which the bomb was cooled from 575°. After the equilibrium temperature had been reached, it was maintained constant for 4 hours before analyzing, temperature readings being taken every 10 minutes.

Valve V_1 was shut tight, and the analytical train described below was connected to the vapor outlet tube. Valve V_2 was then opened very cautiously, and the vapor phase was bled off into the analytical train. This operation caused a good deal of trouble, as it was difficult to open the valve slowly enough to prevent the vapor phase, at a temperature of 500° and a pressure of several hundred atmospheres, from blowing

through the analytical train or from blowing out the stoppers of the absorption tubes. The difficulty was overcome by the use of a worm gear wheel on the stem of valve V_2. This wheel was operated by a worm that turned on a shaft held by a post (not shown in Figure 1) screwed into the valve block. With this arrangement, it was possible to turn the valve stem very slowly, and, by observing due care, to allow the vapor phase to escape very slowly into the analytical train. After its adoption, only a few runs were lost because of too rapid release of the vapor phase.

When practically all the vapor phase had been bled off, as judged from the temperature of the outlet tube, the analytical train and the vapor side of the apparatus were evacuated to 50 mm. The absorption tubes were filled with dry, CO_2-free air and were weighed. The heating units and thermocouples were disconnected, and the apparatus was removed from the furnace and cooled rapidly by standing it in 2 inches of water while air was blown on it by a fan. The bombs were disconnected from the valve block as soon as they were sufficiently cool to be handled. If the melt had been quenched successfully, the crucible and its contents were weighed as a rough check on the analytical results, the volume of the melt was determined, and the melt was examined under the petrographic microscope.

The vapor phase was bled off into the analytical train, which consisted of a copper water-trap, followed by an absorption tube filled with Anhydrone and three absorption tubes containing Ascarite backed by Anhydrone. The copper water-trap was connected to the vapor outlet tube by means of a cone fitting and to the absorption tube filled with Anhydrone by heavy-walled rubber tubing. It was immersed in ice up to the bend in its outlet tube. This simple device was very effective. Even when as much as 4 grams of water was present in the vapor phase, the amount not condensed in the trap and absorbed by the Anhydrone rarely exceeded 2 milligrams. The three Ascarite tubes served to absorb the carbon dioxide. It was necessary to fill the first of these tubes with approximately equal volumes of Ascarite and glass beads to prevent plugging, because the gas reaching this point was nearly pure carbon dioxide. Usually, unless the vapor phase was released into the analytical train too rapidly, not more than 5 per cent of the CO_2 passed through the first tube, and practically none through the second.

It had been thought that an appreciable amount of carbon dioxide would dissolve in the water condensed in the copper water-trap. Numerous analyses of this water were made, and it was found that the amount of CO_2 dissolved was negligible, probably because the trap was designed so that the vapor would not bubble through condensed water and because

the evacuation of the train removed most of the CO_2 that had dissolved. It was also found that evacuating the absorption tubes and refilling them with dry, CO_2-free air caused no change in weight, so that no error was introduced by this procedure.

The accuracy of the analysis for water was checked by weighing known quantities of water into the apparatus, heating to 500°, and analyzing. Three such tests gave a maximum error of —0.8 per cent and an average error of —0.45 per cent.

CALCULATION OF RESULTS

Since a portion of the vapor phase is in bomb A_1 when valve V_1 is closed, the analytical results must be multiplied by a factor to give the total vapor phase. The factor is the ratio of the total vapor volume within the apparatus to the volume occupied by that part of the vapor which is withdrawn for analysis. The volumes of the inner chambers of the bombs were determined by weighing them filled with water and they were also calculated from the dimensions. The total volume of the system, including the bomb chambers, the plunger openings, and the valve block openings was found to be 24.58 cc., of which 21.14 cc. was the volume occupied by the analyzed portion of the vapor—*i.e.*, the volume to the right of valve V_1 in Figure 1. During an actual run, however, the total volume of the system is reduced by the volume occupied by the liquid phase. This could not be determined at the temperature of the experiment, but, when the liquid phase could be quenched successfully, its volume could be determined after the crucible was removed from the bomb. A number of such determinations were made for experiments in which the ratio $K_2O : SiO_2$ was 1 : 3 or 1 : 4, and the volume of the liquid phase varied from 0.75 cc. to 0.88 cc. When the ratio $K_2O : SiO_2$ was 1 : 1, the liquid phase could not be quenched successfully, and it was not possible to determine its volume accurately. A number of approximate determinations were made, and these gave for the volume of the liquid phase values ranging from 0.93 cc. to 1.05 cc. These are believed to be somewhat high, since the melt had frothed a little during quenching.

However, the correction factor is not greatly affected by small changes in the volume of the liquid phase. With a liquid phase of 0.75 cc. volume, the correction factor is calculated to be 1.127; with a liquid phase of 1.05 cc., the correction factor is calculated to be 1.114. The factor was also determined directly by placing 0.85 cc. of copper shot in the platinum crucible in bomb A_1, weighing in water, heating to 500°, closing valve V_1, and determining the amount of water on each side of the valve. Three such determinations gave for the factor 1.117, 1.125, and 1.127, in

good agreement with the calculated value. Since the change in the correction factor caused by the change in the volume of the liquid phase is within the error of the determination of the value of the factor, 1.120 was used for all the calculations.

TABLE 1.—*Log of a typical experiment*

	A. Charge (grams)	B. Analysis of vapor (grams)	C. Total vapor phase (B×1.12) (grams)	D. Liquid phase (A−C) (grams)	Remarks
K_2O.....	0.592	.....	.....	0.592	Weight of quenched liquid phase = 2.043 g. Liquid phase a homogeneous glass, $n = 1.496$.
SiO_2.....	1.132	.....	.....	1.132	
CO_2.....	0.277	0.1996	0.2236	0.053	
H_2O.....	3.949	3.277	3.670	0.279	
				2.056	

The data obtained in a typical experiment, No. 48, of Table 5 are given in Table 1 to illustrate the calculations. The charge, column A, was actually weighed as 0.8687 g. K_2CO_3 and 1.1317 g. SiO_2. The amount of water has been corrected for 0.002 g. lost during evacuation. Column B gives the analysis after subtraction of the 0.0015 g. blank for the CO_2 determination. It will be noted in the column headed *Remarks* that the quench was successful and that the weight of the quenched liquid phase is in excellent agreement with the calculated weight of the liquid phase.

It should be noted that, in the example given in Table 1, an error in the analysis of the vapor phase causes a much larger relative error in the amount of water in the liquid phase. If, for instance, we assume an error of 30 milligrams, or somewhat less than 1 per cent, in the analysis of the vapor phase for water, there is likewise an error of 30 milligrams in the composition of the liquid phase, but here the relative error is nearly 11 per cent. This relative error varies according to the amount of water dissolved in the liquid phase, the range being from 5 to 12 times the percentage error in the analysis.

The over-all reproducibility of the results is in accord with the results of the tests, which indicated an accuracy of better than 1 per cent in the analysis of the vapor phase. The results obtained in duplicate experiments are presented under attainment of equilibrium. The agreement is better than 1 per cent in all cases for the composition of the vapor phase.

ATTAINMENT OF EQUILIBRIUM

It was important to ascertain whether the results obtained were true equilibrium results. Duplicate experiments were carried out, with varying time allowed for reaching equilibrium, the equilibrium temperature being approached from both above and below. The results obtained in several such trials are given in Table 2. They show conclusively that equilibrium was reached in our experiments, the final results being unaffected by approach from higher or lower temperatures or by variations in the materials used to make up the charge.

TABLE 2.—*Attainment of equilibrium under differing conditions*

Experiment number	Ratio SiO_2 / K_2O	Charge	Time hours*	Temperature approached from	H_2O in charge (grams)	H_2O in vapor (grams)	CO_2 in vapor (grams)
11	3	$KHCO_3$ + SiO_2	54	1° above	2.085	1.874	0.420
12	3	$KHCO_3$ + SiO_2	27	5° above	2.078	1.880	0.405
14	3	$KHCO_3$ + SiO_2	23	Below	2.089	1.885	0.406
38	3	K_2CO_3 + SiO_2	21	Below	2.541	2.288	0.2195
55	3	K_2CO_3 + SiO_2	22	2° above	2.557	2.303	0.224
48	3	K_2CO_3 + SiO_2	19	Below	3.949	3.670	0.224
53	3	K_2CO_3 + SiO_2	23	2° above	3.947	3.636	0.220
83	1	K_2CO_3 + SiO_2	23	Below	2.360	1.985	0.154
84	1	$KHCO_3$ + $K_2Si_2O_5$	29	70° above	2.315	1.946	0.153

* From start of heating.

RATE OF REACTION OF QUARTZ WITH CARBONATE SOLUTIONS

It was desired to use SiO_2 in making up the charges so as to get the maximum amount of CO_2, but it was not known if quartz would dissolve in a carbonate solution rapidly enough to complete the run in a reasonable time. Accordingly a series of experiments was made on the rate of solution of quartz in carbonate solutions at different temperatures and for varying periods of time. In these experiments, quartz (20-40 mesh) and $KHCO_3$ in the ratio SiO_2 / K_2O = 3 were placed in a platinum crucible in a bomb of 40 cc. capacity, water was added, and the bomb was heated to the desired temperature. The time required for complete solution of the quartz at a given temperature was determined by making a series of runs of varying duration, quenching, and examining the charge under the

petrographic microscope. The results are given in Table 3 and are shown in Figure 3.

The results above are a summary of some 80 runs. The time is that from the start of heating. A Variac in the furnace line was set so that

TABLE 3.—*Reaction of quartz with carbonate solutions*

Temp. °C.	Time* hours	H_2O added g.	Solid phase
340	>220	3.0	$K_2Si_2O_5$
355	190	3.0	$K_2Si_4O_9.H_2O$
365	120	3.0	$K_2Si_4O_9.H_2O$
374	90	3.0	$K_2Si_4O_9.H_2O$
382	58	3.0	$K_2Si_4O_9.H_2O$
404	36	3.0	$K_2Si_4O_9.H_2O$
424	26	3.0	$K_2Si_4O_9.H_2O$
441	23	3.0	$K_2Si_4O_9.H_2O$
452	20	3.0	$K_2Si_4O_9.H_2O$
468	16	3.0	Glass (unsaturated solution)
400	32	10.2	Glass (unsaturated solution)
424	20	6.4	Glass (unsaturated solution)

* Time (from start of heating) required for complete solution of quartz.

the current would hold the desired temperature, and the furnace was heated slowly to avoid spattering. About 12 hours was required before the final temperature was reached. At the higher temperatures, therefore, much of the quartz reacted before the final temperature was reached, and the results given are not very close approximations to the time that would be required at the final temperature.

The rapid increase in the time required for solution of the quartz as the temperature is decreased is due to a number of factors, among which are decreasing temperature, decreasing pressure, and decreasing solubility of the potassium silicates. At this particular ratio of $SiO_2 : K_2O$, the solubility decreases very rapidly below 450°, and this is probably the chief factor causing the rapid increase in the time required for solution of the quartz. In the experiments below 400°, the quartz was nearly always found as grains surrounded by the sheath of $K_2Si_2O_5$, which undoubtedly slowed down the reaction greatly. The broken line in Figure 3 represents the rate of solution of quartz when sufficient water was present to dissolve the potassium silicate formed. Presumably, the two curves would diverge even more at temperatures below 400°, but this could not be tested experimentally, as the solubility is so low that sufficient water could not be added to the charge in the bomb.

Some runs were made with quartz of approximately 100 mesh, but the rate was only slightly more rapid than when 20-40 mesh material was used.

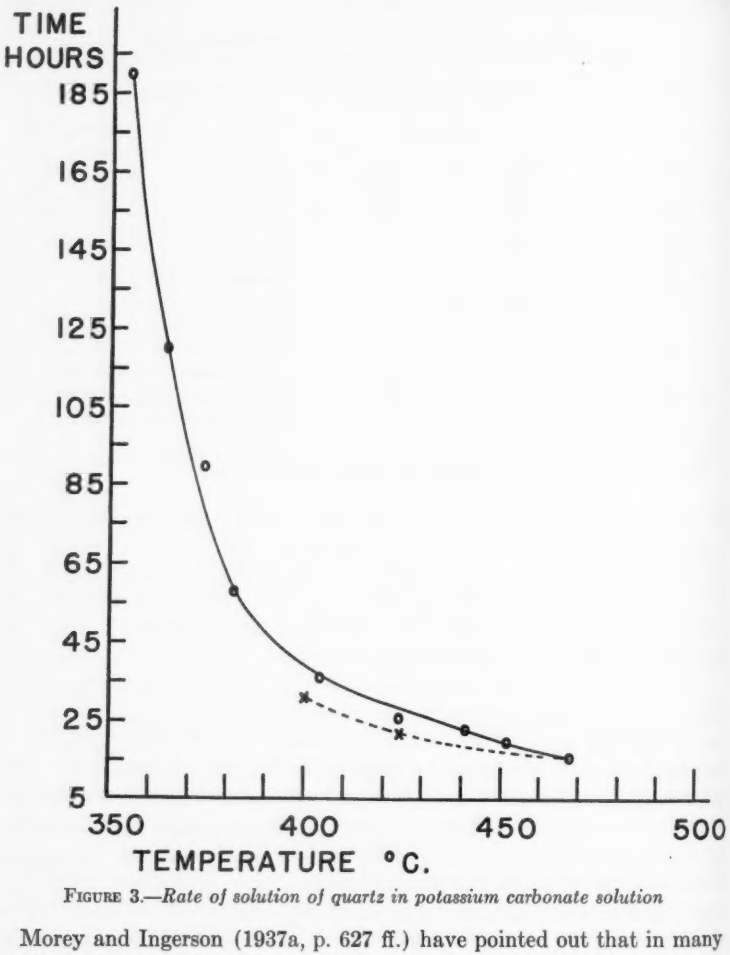

FIGURE 3.—*Rate of solution of quartz in potassium carbonate solution*

Morey and Ingerson (1937a, p. 627 ff.) have pointed out that in many of the hydrothermal syntheses of silicates reported in the literature the products obtained undoubtedly did not represent true equilibrium products. The use of material that reacts extremely slowly has the effect of changing the composition, as part of the material is not effective in the

reactions that occur. Examples of this have been observed repeatedly. Table 4 is a typical case. In this case, the composition is close to that at which the disilicate is stable, and it is only when practically all the quartz has dissolved that the hydrated tetrasilicate becomes the stable

TABLE 4.—*Change of solid phase as equilibrium is approached*

Charge—$KHCO_3$ 1.105 g., SiO_2 0.995 g., H_2O 3.0 g., $SiO_2/K_2O = 3$

Time of run, hours	Temperature °C.	Solid phases observed in charge
81	356	Much quartz, much $K_2Si_2O_5$
95	355	Moderate quartz, much $K_2Si_2O_5$
138	355	Little quartz, much $K_2Si_2O_5$, very little $K_2Si_4O_9 . H_2O$
188	355	Trace quartz, no $K_2Si_2O_5$, much $K_2Si_4O_9 . H_2O$

phase. If the experiment had been run no longer than 4 days, it might have been erroneously assumed that the disilicate is the stable solid phase at this temperature and ratio of K_2O to SiO_2. The importance of using homogeneous starting material and of making sure that equilibrium is reached cannot be stressed too strongly.

DISCUSSION OF RESULTS

The experimental results are assembled in Table 5 and shown in Figures 3 to 7. It is difficult to present these results so as to give a picture that can be readily grasped. The system is one of four components, in two phases, and at constant temperature, which leaves three degrees of freedom. The experiments may be considered in four groups, which have ratios of K_2O to SiO_2 of 1 : 1, 1 : 2, 1 : 3, and 1 : 4. Since in no case is the partial pressure of CO_2 great enough to precipitate a carbonate, in each of these groups the degrees of freedom can be considered as reduced to two.

Consider, then, the system CO_2—H_2O—K_2SiO_3 at 500° C. There are three components, in two phases at constant temperature, thus two degrees of freedom. Fixing the pressure does not define the system because the partial pressures of CO_2 and H_2O are still variable. A given liquid composition has a uniquely determined vapor composition; but in fixing the liquid composition two additional restrictions have been imposed. With a given content of CO_2 there can be a series of solutions of variable content of water and variable vapor phase. Fixing the vapor composition

leaves one degree of freedom, and a vapor of a given percentage composition in CO_2 and H_2O may be given off by a series of solutions. The system does, however, become invariant (with constant ratio of K_2O : SiO_2) when in addition to the vapor composition the total pressure is fixed.

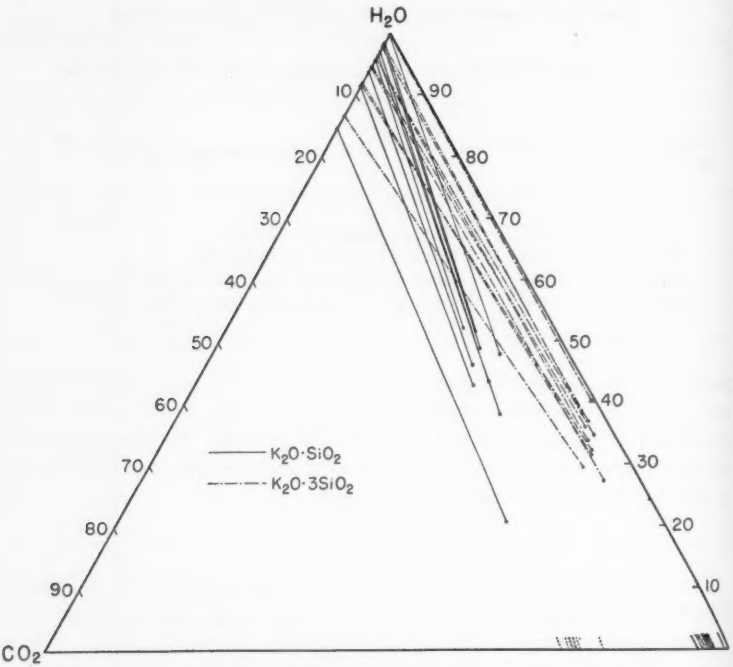

FIGURE 4.—*Compositions of some coexisting vapor and liquid phases*

Solid lines connect vapor and liquid in metasilicate melts; dot-dash lines, in trisilicate melts

This is equivalent to fixing the partial pressures of the two constituents of the vapor.

The experimental study has deliberately been restricted to solutions above the liquidus. In the ternary system CO_2—H_2O—K_2SiO_3, the appearance of a solid phase reduces the degrees of freedom to one. Crystalline potassium metasilicate at 500° can coexist in equilibrium with a series of liquids, of variable ratio of CO_2 : H_2O. Fixing any additional variable, such as the total pressure, or the ratio CO_2 : H_2O in either liquid or vapor, defines the system completely. If the CO_2 content be increased to the point at which carbonate separates, the system CO_2—H_2O—K_2SiO_3

TABLE 5.—*Experimental results*

$K_2O:SiO_2=1:1$ 500°

Charge—$KHCO_3$ 1.653 g., SiO_2 0.496 g.=K_2O 0.778 g., SiO_2 0.496 g., CO_2 0.727 g.

Run (number)	Time of run (hours)	Equilibrium approached from	H_2O in charge (grams)	Liquid phase				Vapor phase		% of total CO_2 in vapor	Partial pressure H_2O atmosphere	Partial pressure CO_2 atmosphere
				K_2O (grams)	SiO_2 (grams)	CO_2 (grams)	H_2O (grams)	CO_2 (grams)	H_2O (grams)			
70*	22	10° above	1.086	0.778	0.496	0.292	0.110	0.435	0.976	59.8	141	26
67	24	2° above	2.169	0.778	0.496	0.295	0.313	0.432	1.856	59.4	250	26
66	22	Below	3.166	0.778	0.496	0.282	0.383	0.444	2.783	61.3	326	27
69	23	Below	4.181	0.778	0.496	0.288	0.459	0.439	3.722	60.5	385	27

$K_2O:SiO_2=1:1$ 500°

Charge—K_2CO_3 1.394 g., SiO_2 0.606 g.=K_2O 0.950 g., SiO_2 0.606 g., CO_2 0.444 g.

71	26	7° above	1.951	0.950	0.606	0.286	0.297	0.158	1.654	35.5	202	9.6
84†	29	70° above	2.315	0.950	0.606	0.291	0.369	0.153	1.946	34.5	230	9.3
83	23	Below	2.360	0.950	0.606	0.290	0.375	0.154	1.985	34.8	233	9.4
72	24	Below	3.109	0.950	0.606	0.291	0.466	0.153	2.643	34.5	288	9.3
87	24	Below	4.094	0.950	0.606	0.286	0.516	0.158	3.578	35.5	353	9.6

$K_2O:SiO_2=1:1$ 500°

Charge—K_2CO_3 0.784 g., $K_2Si_2O_5$ 1.216 g.=K_2O 1.069 g., SiO_2 0.682 g., CO_2 0.250 g.

88	28	Below	2.549	1.069	0.682	0.241	0.444	0.009	2.105	3.5	242	0.5
86	24	Below	3.052	1.069	0.682	0.240	0.461	0.010	2.591	6.3	285	0.6
90	27	Below	4.034	1.069	0.682	0.239	0.471	0.011	3.563	7.0	353	0.7

88	28	Below	2.549	1.069	0.682	0.241	0.444	0.009	2.105	3.5	242	0.5
86	24	Below	3.052	1.069	0.682	0.240	0.461	0.010	2.591	6.3	285	0.6
90	27	Below	4.034	1.069	0.682	0.239	0.471	0.011	3.563	7.0	353	0.7

$K_2O:SiO_2=1:2$ 500°

Charge—$KHCO_3$ 1.325 g., SiO_2 0.795 g.=K_2O 0.623 g., SiO_2 0.795 g., CO_2 0.582 g.

75	23	Below	3.129	0.623	0.795	0.125	0.327	0.457	2.802	78.5	325	28

$K_2O:SiO_2=1:3$ 500°

Charge—$KHCO_3$ 1.105 g., SiO_2 0.994 g.=K_2O 0.520 g., SiO_2 0.994 g., CO_2 0.486 g.

23	21	Below	1.177	0.520	0.994	0.113	0.187	0.373	0.990	76.8	130	23
12	27	5° above	2.078	0.520	0.994	0.081	0.198	0.405	1.880	83.4	224	25
11	54	1° above	2.085	0.520	0.994	0.066	0.211	0.420	1.874	86.5	223	26
14	23	Below	2.089	0.520	0.994	0.080	0.204	0.406	1.885	83.6	225	25
27	21	8° above	3.141	0.520	0.994	0.061	0.238	0.425	2.903	87.5	308	26
63	26	Below	3.238	0.520	0.994	0.066	0.250	0.420	2.988	86.5	314	26
65	25	Below	4.212	0.520	0.994	0.079	0.294	0.407	3.918	83.8	374	25

$K_2O:SiO_2=1:3$ 500°

Charge—K_2CO_3 0.869 g., SiO_2 1.132 g.=K_2O 0.592 g., SiO_2 1.132 g., CO_2 0.277 g.

45‡	19	Below	1.041	0.592	1.132	0.125	0.071	0.152	0.970	54.9	127	9.2
57	23	Below	1.498	0.592	1.132	0.093	0.183	0.183	1.315	66.3	167	11
52	26	2° above	1.500	0.592	1.132	0.088	0.162	0.189	1.338	68.3	169	11
38	21	Below	2.541	0.592	1.132	0.057	0.253	0.220	2.288	79.4	261	13
55	22	2° above	2.557	0.592	1.132	0.053	0.254	0.224	2.303	80.9	261	14
53	23	2° above	3.947	0.592	1.132	0.056	0.311	0.220	3.636	79.7	354	13

52	26	2° above	1.500	0.592	1.132	0.088	0.162	0.189	1.338	68.3	169	11
38	21	Below	2.541	0.592	1.132	0.057	0.253	0.220	2.288	79.4	261	13
55	22	2° above	2.557	0.592	1.132	0.053	0.254	0.224	2.303	80.9	261	14
53	23	2° above	3.947	0.592	1.132	0.056	0.311	0.220	3.636	79.7	354	13
48	19	Below	3.949	0.592	1.132	0.053	0.279	0.224	3.679	80.9	359	14
61	23	Below	5.128	0.592	1.132	0.063	0.326	0.213	4.802	77.1	410	13

$K_2O:SiO_2 = 1:3$ 500°

Charge—K_2CO_3 0.243 g., $K_2Si_4O_9$ 1.758 g. = K_2O 0.660 g., SiO_2 1.263 g., CO_2 0.077 g.

96	43	Below	1.548	0.660	1.263	0.015	0.235	0.062	1.313	80.0	167	4
92	25	8° above	3.267	0.660	1.263	0.014	0.340	0.064	3.267	82.4	334	4

$K_2O:SiO_2 = 1:4$ 500°

Charge—$KHCO_3$ 0.948 g., SiO_2 1.137 g. = K_2O 0.446 g., SiO_2 1.137 g., CO_2 0.417 g.

98	23	Below	3.497	0.446	1.137	0.064	0.246	0.353	3.251	84.7	332	21
100	45	[illegible] above	3.643	0.446	1.137	0.067	0.259	0.350	3.384	84.0	341	21
97	4[illegible]	Below	2.569	0.446	1.137	0.036	0.312	0.381	2.257	91.5	259	23

$K_2O:SiO_2 = 1:4$ 500°

Charge—K_2CO_3 0.730 g., SiO_2 1.270 g. = K_2O 0.498 g., SiO_2 1.270 g., CO_2 0.233 g.

9[illegible]	26	2° above	2.517	0.498	1.270	0.029	0.188	0.204	2.329	87.7	263	12

* Probably below liquidus.
† Charge same composition as No. 83 but made up from $KHCO_3$ + $K_2Si_2O_5$.
‡ Below liquidus, crystals of $K_2Si_2O_5$ in quenched charge.

can no longer be considered as ternary; the point obtained is on the univariant equilibrium V—L—S_1—S_2 in the quaternary system, but the composition of the liquid can no longer be expressed in terms of these three components.

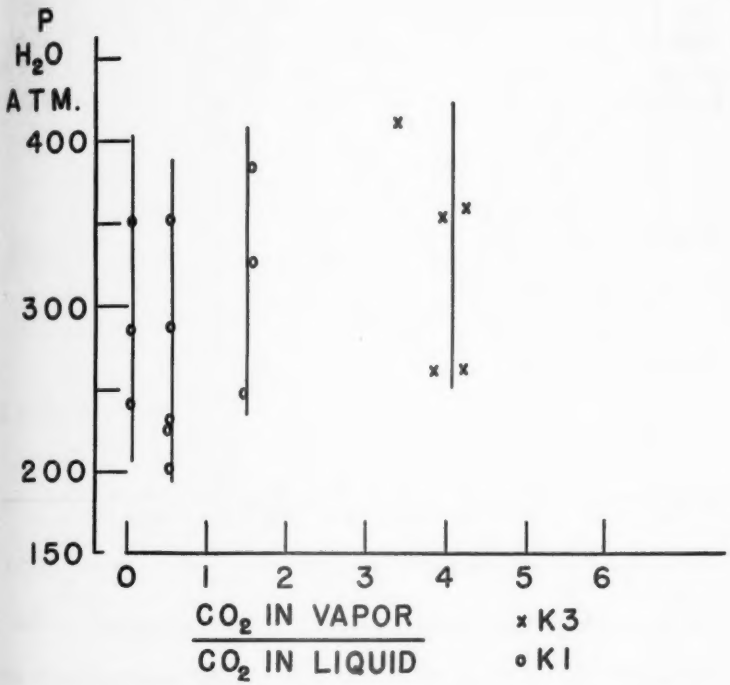

FIGURE 5.—*Distribution of CO_2 between vapor and liquid with increasing pressure of H_2O*

In Figure 4 the coexisting vapor and liquid compositions for the $K_2O : SiO_2$ ratios 1 : 1 and 1 : 3, calculated in mol percentage, are connected by tie-lines in a ternary diagram. This diagram may be regarded as a projection of the tetrahedron representing the system CO_2—H_2O—K_2O—SiO_2 on the side CO_2—H_2O—K_2O, with the eye-point at the SiO_2 apex. It will be observed that the same vapor composition may be obtained with a large number of liquid compositions.

The direction or slope of the tie-lines is a measure of the distribution ratio of $CO_2 : H_2O$ between vapor and liquid phase. On tie-lines which

(or the prolongation of which) pass through the apex opposite the side CO_2—H_2O, the ratio (mol percentage of CO_2 in vapor : percentage of CO_2 in liquid) is unity; on a tie-line which intersects the base opposite the H_2O apex, that ratio is less than unity, and relatively more water than

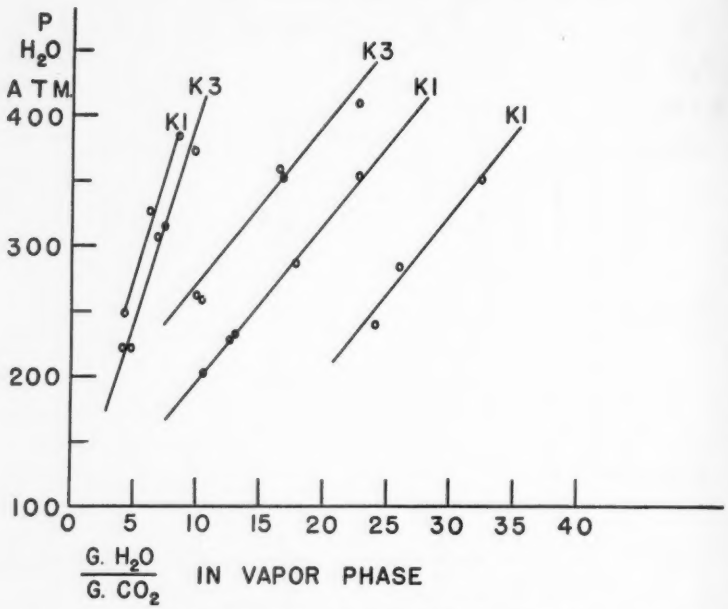

FIGURE 6.—*Change in composition of the vapor with increasing pressure of H_2O*

CO_2 passes into the vapor phase; and if the tie-line passes through the side opposite the CO_2 apex, the ratio of CO_2 to water is greater in the vapor than in the liquid.

With metasilicate mixtures, the distribution ratio is strongly toward retention of the CO_2 in the liquid phase. With increasing content of CO_2 the ratio increases—that is, a greater proportion of CO_2 passes into the vapor.

In each series of runs the total CO_2 content is constant. For example, in the first series of metasilicate ratio, made from $KHCO_3$ and SiO_2, the total amount of CO_2 present is 0.737 g.; in the series made from K_2CO_3 and SiO_2, 0.444 g.; and in the series made from K_2CO_3 and $K_2Si_2O_5$, 0.250 g. The actual amount of CO_2 present in the vapor is approximately constant in each of the series, although the water content

is greatly increased within each series. (*See* Figure 5.) It follows, then, that the percentage of CO_2 in the vapor decreases as the partial pressure of H_2O is increased. This is shown by the curve of Figure 6. This must be the case if the distribution ratio is to remain constant. Therefore, if

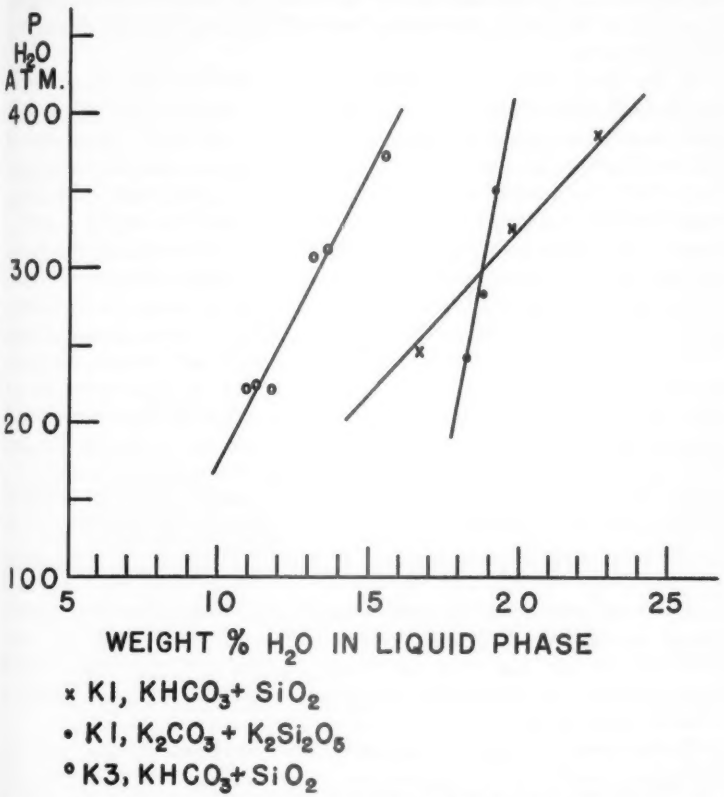

FIGURE 7.—*Distribution of water between vapor and liquid at constant temperature and varying pressure*

the pressure is reduced by escape of vapor, the composition of the vapor phase will change with drop in pressure, in the direction of increase in the percentage of CO_2.

In the preceding discussion, attention has been centered on the vapor phase. Consider the first two series of experiments of metasilicate ratio, made from $KHCO_3$ and quartz, and K_2CO_3 and quartz, respectively,

having average contents of CO_2 in the liquid of 0.292 and 0.288. These are identical within the limit of error of the method. The compositions of these melts can be calculated in terms of the molecules K_2SiO_3, $K_2Si_2O_5$, and K_2CO_3, and it is evident that the proportions of these three molecules are not affected by their varying water content, ranging from 6.56 to 22.7, or by the change in partial pressure of CO_2 from 9 to 27 atmospheres.

In the third series of experiments of metasilicate ratio, made from K_2CO_3 and $K_2Si_2O_5$, the total amount of CO_2 present is 0.250 g., less than that in the liquid phase in the preceding two series. Now almost all the CO_2 is in the liquid. Apparently, as CO_2 is added to the system H_2O—$K_2Si_2O_5$, most of it remains in the liquid phase until a definite proportion of carbonate is formed, after which the CO_2 remains in the vapor. Up to this limiting value, CO_2 is rapidly taken up by the melt; beyond this limiting value it is less soluble in the melt than water vapor.

The partial pressure of water was calculated by means of the tables given by Keenan and Keyes (1936), that of CO_2, from the ideal gas law, with the assumption of conformity with Dalton's Law in each case. It is probable that this assumption is in error, but, because there are no data on the behavior of CO_2—H_2O mixtures at temperatures and pressures close to those in the present experiments, it seems best to use the simplest assumptions. When such data are available, the partial pressures can easily be recalculated, since the volumes and masses are given in Table 5.

The distribution ratio of CO_2 in the vapor to that in the liquid increases as the ratio of $SiO_2 : K_2O$ increases from metasilicate to tri- and tetrasilicate. With the trisilicate liquids the ratio is a little less than one; with tetrasilicate liquids it appears to be about one; but more experiments are needed. Also, with increasing ratio of $SiO_2 : K_2O$, the solubility of CO_2 and H_2O each decreases, but that of CO_2 decreases much more rapidly. It is probable that these conclusions will also apply to natural magmas.

The amount of water in the liquid phase increases linearly with the pressure of water in the vapor phase, in accordance with Henry's Law, as was previously found by Morey and Fenner (1917) in the system H_2O—K_2SiO_3—SiO_2. This is shown in Figure 7, in which the percentage by weight of water is plotted against pressure of water in atmospheres for three different mixtures. In one of these mixtures the ratio of $K_2O : SiO_2$ was 1 : 3; in the other two, 1 : 1. The two metasilicate mixtures were not of the same composition. In one series the ingredients were $KHCO_3$ and SiO_2; in the other, K_2CO_3 and $K_2Si_2O_5$.

The study presented in this report represents little more than the working out of a method, and continuation of the work is planned. It is

desirable to follow each of the composition mixtures down to the appearance of crystals and to increase the pressure of CO_2 until crystalline carbonate appears. The writers have planned to complete this isotherm at 500° before extending the study to other temperatures and prefer not to extrapolate the experimental results to their geological applications until this further work has been completed.

SUMMARY

A study has been made of the equilibrium between vapor and liquid in the system CO_2—H_2O—K_2O—SiO_2, at 500°. Partial pressures of H_2O ranged up to 400 atmospheres, of CO_2, up to 25 atmospheres, and the mixtures used had ratios of K_2O : SiO_2 from 1 : 1 to 1 : 4.

The distribution ratios of the fraction of CO_2 in the vapor to the ratio of the percentages of CO_2 to H_2O in the liquid are indicative of the tendency of the metasilicate melts to retain CO_2 and of the tendency toward expulsion of CO_2 with more siliceous mixtures. With drop in pressure at constant temperature and ratio K_2O : SiO_2, the composition of the vapor changes toward higher contents of CO_2.

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DIFFERENTIATION OF THE PALISADE DIABASE, NEW JERSEY

BY FREDERICK WALKER

CONTENTS

	Page
Abstract	1060
Introduction	1060
Acknowledgments	1061
Field characters	1063
Introduction	1063
Olivine diabase layer	1064
Normal diabase	1065
Residual injections	1065
Allied intrusions	1066
Contact alteration	1067
Mineralogy	1067
General considerations	1067
Olivine	1067
Plagioclase	1068
Pyroxene	1070
General statement	1070
Hypersthene-augite	1070
Pigeonite	1071
Hypersthene	1072
Clino-hypersthene (?)	1073
Relative abundance and distribution of ortho- and clinopyroxene	1073
Composition of pyroxene	1073
Minor constituents	1074
Iron ore	1074
Biotite	1074
Micropegmatite	1075
Amphibole	1076
Accessories	1076
Micrometric data	1077
Accuracy	1077
Data	1078
Chemical data	1078
Petrology	1079
History of investigation	1079
The case of gravitational differentiation	1083
Olivine	1083
Pyroxene	1087
Crystallization of the pyroxenes	1089
Later stages of differentiation	1092

Page

The case against syntexis 1096
Chemical analyses 1098
Differentiation of similar intrusions 1099
General conclusions 1101
Works to which reference is made 1101

ILLUSTRATIONS

Figure Page

1. Map of Palisade sill 1062
2. Diagrammatic section across the Palisade sill 1066
3. Average size of the olivine and plagioclase 1069
4. Anorthite content of plagioclase in the sill 1070
5. Mineral composition of the lower part of the section 1074
6. Amount of plagioclase and pyroxene and size of plagioclase 1075
7. Specific gravities of sections 1089
8. Triangular diagram of pyroxenes of Palisade sill 1091
9. MgO and FeO + Fe_2O_3 plotted against height above lower contact of Palisade sill 1099

Plate Facing page

1. Views of the Palisade sill 1104
2. Photomicrographs of rocks of the Palisade sill 1105

ABSTRACT

The Palisade diabase sill of New Jersey has been widely cited as a type case of gravitational differentiation, and the present investigation is an attempt to explain certain inconsistencies which appear in previous accounts and to make them completely convincing. New data include 15 chemical analyses and micrometric measurements of 150 rock sections. They show that crystal fractionation can account satisfactorily for nearly all the differentiation phenomena displayed by the sill.

INTRODUCTION

The great Palisade diabase sill of New Jersey has long received the attention of geologists. Its majestic cliff line along the Hudson River, its admirably exposed contacts, and most of all, perhaps, its situation adjacent to the largest city of the globe have drawn many to its study. In some respects, however, their investigations remain incomplete though the problems involved are of prime importance to petrologists.

The only detailed petrological account is that of Lewis (1908a, p. 105-147). It appeared over 30 years ago but was undoubtedly in advance of its time. Full descriptions of the field characters and petrography of the intrusion were supplemented by considerable chemical and micrometric data. A very clear account was given of its differentiation; Lewis argued that the layer of olivine diabase, which occurs about 50 feet above the lower contact, was due to the gravitational settling of olivine crystals *in situ*. The sill has ever since been cited as one of the best known

examples of this process. There were, however, several discrepancies in Lewis' data upon which later authors have commented, though no systematic attempt seems to have been made to explain them satisfactorily. The present investigation is an attempt to find the cause of these discrepancies by means of a detailed examination of critical sections and the production of more complete and connected chemical and micrometric data. Brief visits were made to the Palisades in 1934 and 1936, and a systematic collection along the best exposed portions was carried out in the winter of 1937.

The Palisade diabase seems to the author to demand specially thorough quantitative examination for the following reasons:

(1) The magma of the sill in its undifferentiated state is representative of a type of world-wide distribution, *viz.*, the tholeiite magma type of Wahl (1908, p. 69) and Kennedy (1933, p. 240-242).

(2) The thickness of the sill was sufficient to allow slow cooling and a very considerable degree of differentiation, yet it is not too great to prevent relatively rapid traverses being made from top to bottom.

(3) The fact that the intrusion possesses both roof and floor has resulted in the accumulation of the lighter and heavier products of differentiation in positions where they may be seen.

(4) The situation of the outcrop in or adjacent to a densely populated region is responsible for unrivaled exposures, principally along new highways blasted out of solid rock.

(5) Unusually accurate data are available as to the physical properties of the magma.

These factors render the intrusion unique in the advantages which it affords to investigators.

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The laboratory work in connection with this investigation was carried out in four different universities: Columbia, Harvard, Cambridge, and Cape Town, and the author is greatly indebted to the heads of the various petrological departments which afforded him hospitality—*i.e.*, to Professors Douglas Johnson and S. J. Shand of Columbia, to Professor E. S. Larsen of Harvard, to Professor C. E. Tilley of Cambridge. It is also a pleasure to record the courtesy and assistance which he invariably received from all members of the staffs of these departments with whom he came in contact. He is particularly grateful to Professor E. S. Larsen, who accompanied him in the field and saw the manuscript through the press, and to Professor R. A. Daly of Harvard for much stimulating discussion.

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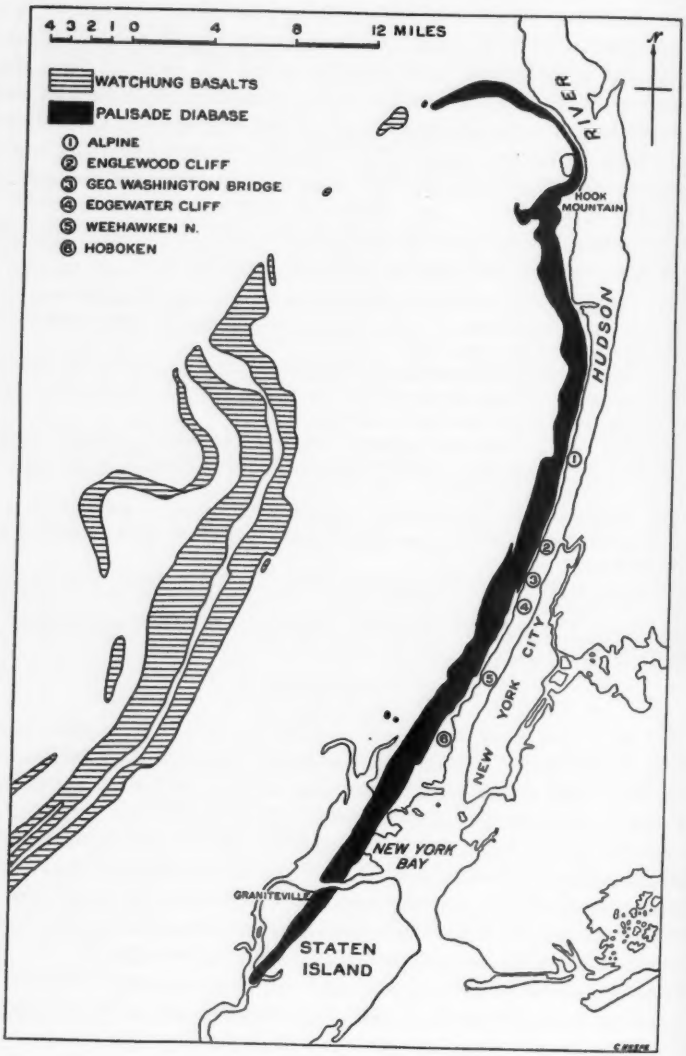


FIGURE 1.—*Map of Palisade sill*

FIELD CHARACTERS

INTRODUCTION

Detailed accounts of the field relations of the Palisade diabase have been given by Darton (1890, p. 37-53), Merrill and others (1902, p. 8-10), Darton and others (1908, p. 11-12), Lewis (1907a; 1908a, p. 105-109), and others, but so much building and road construction has been carried out since the latest of these that some revision is necessary.

The main outcrop of the sill extends from Haverstraw, New York, along the west bank of the Hudson River to Jersey City and from there across the Kill van Kull to Fresh Kills on Staten Island. The olivine diabase layer is, however, exposed only between Jersey City and Alpine. Detailed collection was therefore confined within these limits, though representative specimens were also secured from various points between Hook Mountain and Graniteville (Fig. 1). From Jersey City northward, the sill forms a conspicuous escarpment which, north of Edgewater, is usually crowned by a steep cliff. The great columnar joints of this cliff simulate a stockade and are responsible for the name "Palisades". This part of the sill is probably over 1000 feet thick, but it is thought to thin out toward Jersey City. It is intruded into Triassic arkoses and shales of the Newark formation and has a westward dip of about 15 degrees. West of the cliff line ice erosion has scratched and polished many surfaces. The sill disappears below the Newark formations fringing the marshy alluvial flat of the Hackensack River. There is, however, considerable glacial drift on the dip-slope, and much of the ground not yet utilized is covered by second-growth timber.

The lower contact of the sill is better exposed than the upper. It is seen at intervals all the way from Hoboken to Alpine, but the best and most continuous sections are observed along the construction work of the Hendrik Hudson Drive both north and south of the George Washington Bridge. Excellent transgressive contacts are also visible at King's Point, Weehawken, though the sill on the whole is concordant. Slabs of arkose and shale have been floated up by the diabase in many places, and at Weehawken cliff a large slab of arkose stands vertically in the more slowly cooled portion of the sill far above the lower contact. All contacts are knife-sharp and show chilling of the diabase to a black basalt with a tachylitic selvage.

In most places the upper contact is obscured by drift but it may be observed at the entrance to several railroad tunnels, notably at North Bergen, New Durham, and Haverstraw. At all three localities it is markedly discordant. On a bluff above the intersection of Prospect and Fairview Avenues the upper contact is vertical. Numerous conformable

contacts, however, have been recorded, and, although many of them are no longer visible, there is an excellent one to be seen on the right bank of Flatrock Brook just below the bridge where it is crossed by Highway 4. The suggestion has been made that the intrusion becomes dikelike toward the west, but, as Lewis (1907b, p. 120-121) has pointed out, the visible transgressions may be quite local as at King's Point. Intense alteration of the adjacent formations is seen at all contacts, the arkose being bleached and the shales baked to a flinty hornfels. Full accounts of the contact alteration are given by Kümmel (1897, p. 61-72) and Lewis (1908a, p. 138-147).

On cliffs and quarry faces rude polygonal contraction columns are conspicuous and often extend unbroken through the whole height, including the olivine layer when it is present. Major joint faces are frequent, and cylindrical exfoliation cracks may develop in individual columns. Many of the joint planes show traces of movement.

OLIVINE DIABASE LAYER

From Jersey City to Alpine near the base of the sill is an olivine-rich layer or ledge averaging 15 feet thick. This layer is darker and usually weathers more readily than the normal diabase and is therefore conspicuous on cliff faces, particularly where its crumbly disintegration has produced a ledgelike feature. It is 30 to 60 feet above the lower contact, just over the lower chilled phase of the sill. Where the contact is transgressive and cuts across the bedding planes at King's Point, it is usually paralleled by the olivine layer. The olivine layer is distinctly coarser than the normal diabase. Both its upper and lower surfaces are somewhat undulose. From a short distance they appear to be sharply defined, but closer scrutiny often reveals gradational contacts with the normal diabase, and microscopic evidence indicates gradational contact in all cases.

The crumbly olivine diabase layer is traversed by numerous hard bands which Lewis interpreted as dikelike intrusions of the normal diabase. Those sliced by the author, however, proved to be as rich in olivine as the surrounding rock. It was observed, moreover, particularly at Edgewater Cliff, that these hard bands are developed along the vertical division planes which separate the polygonal columns, and the author attributes them to the hardening action of hydrothermal solutions along these channels. Apart from these dikelike bands a few hard spheroidal kernels may be found in the crumbly matrix and provide the only material from which thin slices can be prepared, yet the minerals of the disintegrating material including the olivine are often fresh, resembling in this respect some of the Hebridean picrites.

Although the olivine diabase layer is generally about 15 feet thick it may reach 20 feet and at Edgewater Cliff it thins out locally to 3 feet. Bailey (1936, p. 4) states that it disappears altogether at one point, but the author was unable to confirm this.

The olivine diabase layer is exposed extensively between Jersey City and Alpine. North of Alpine the lower part of the sill is not visible except near Nyack, and though the olivine diabase layer may be present it is concealed by talus. The olivine diabase layer was not seen at Hook Mountain north of Nyack, though the lower part of the sill is exposed. Horizontal bands of crumbly rock on a quarry face at the end of the road 2 miles north of Nyack proved to be normal diabase and are probably crush zones.

NORMAL DIABASE

The normal rock shows a progressive increase in grain size from the basaltic lower contact to the upper chilled phase. There is indeed very little local variation except in a zone 150-200 feet thick immediately below the upper chilled phase. Here there is a patchy development of pegmatitic schlieren which may reach 2 feet across but which are usually thinner. The grain of these schlieren is often very coarse. Bladelike pyroxene crystals may attain lengths of 3 inches and often show marked curvature. Below this pegmatitic zone really coarse diabase is very rare and where it does occur it grades evenly into the normal variety. A few thin irregular veins of coarse material about half an inch across were observed about 20 feet above the lower contact at the roadside just north of the eastern entrance of the Edgewater-Fairview railroad tunnel and also at the top of Edgewater Cliff and 5 feet below the olivine layer at King's Point.

RESIDUAL INJECTIONS

Both the lower and the upper chilled phases are traversed at numerous localities by white veins up to 3 inches across. These veins, which do not seem to have been recorded previously, are of felsitic appearance and evidently are quartzo-feldspathic. They are somewhat irregular in their course but tend to be vertical rather than horizontal. In only one case has such a vein been traced to the contact, and this proved to be in an absolutely inaccessible position on a smooth vertical rock face under a small waterfall. The locality is just above the Hendrik Hudson Drive about a mile north of the George Washington Bridge, where white veins are particularly abundant in the lower chilled phase. They were also found at Englewood Cliff and near the base of the Weehawken section. In the upper chilled phase exactly similar veins were found just west of the summit of the ridge between Fairview Avenue and Dallytown Road.

The mutual relationship of the various types is shown diagrammatically on Figure 2.

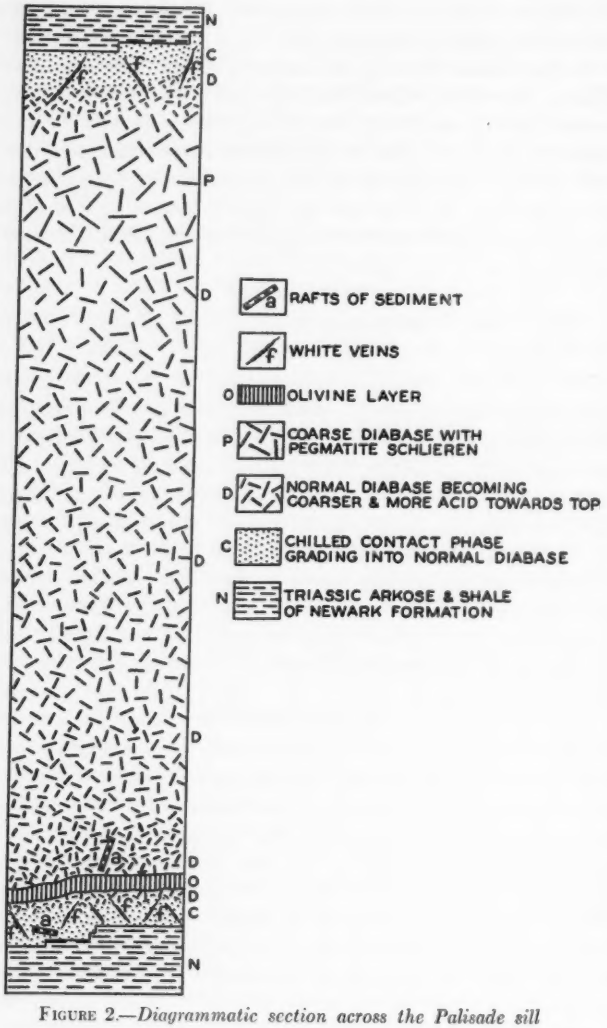


FIGURE 2.—*Diagrammatic section across the Palisade sill*

ALLIED INTRUSIONS

The Palisade diabase is by far the largest basic intrusion in the Newark formation, but other similar basic bodies occur between it and the

Watchung basalt series 8000 feet higher (Lewis, 1907b, p. 122). The most important of these is the 80-foot sill so well exposed in the quarries north of Granton. There can be very little doubt that this intrusion and also the small 18-inch sill intruded into the underlying strata below Coytesville on the Hendrik Hudson Drive are directly connected with the Palisade sheet. All three probably belong to the same period of activity as the Watchung flows and the intervening intrusive masses. Other small sills and veins have a visible connection with the Palisade sheet.

CONTACT ALTERATION

Owing to the great thickness of the Palisade diabase there is considerable contact alteration of the adjacent formations which has been fully described by Lewis (1908a, p. 138-146). Only such metamorphic phenomena as have a direct bearing on the problem of differentiation will be described in this paper.

MINERALOGY

GENERAL CONSIDERATIONS

The Palisade diabase is a typical representative of the tholeiitic magma type, and its mineralogy is therefore simple. Pyroxene and calcic plagioclase are the two main constituents, but olivine occurs locally in considerable amounts, as do quartz and alkalic feldspar, which bear an antipathetic relationship toward it. Accessory minerals include iron ore and biotite. The pyroxenes in the diabase show remarkable variety both in optical properties and chemical composition.

OLIVINE

Only one-eighth of the total thickness of the sill is olivine bearing. The mineral occurs in the upper and lower chilled phases, each of which is 15 to 30 feet thick, in a proportion of about 1 per cent, though as much as 3 per cent was recorded locally. Above the lower chilled phase the proportion increases steadily up to within a foot or two of the olivine diabase layer. Above this level it increases much more rapidly and reaches a maximum of 25 per cent in the olivine layer. An equally rapid decrease occurs just above the olivine layer, and the mineral disappears altogether a few feet from the top of that layer.

Olivine seems to have been the first constituent to crystallize and is idiomorphic except when resorbed. The micrometric data show that in the chilled phases the average diameter of the grains is 0.2 mm.+. There is some variation in size, however, and the maximum diameter is 0.5 mm. The mineral frequently forms unorientated groups and generally shows marked resorption. In some cases it is, as Lewis (1908a, p. 128) has observed, surrounded by a reaction rim of small pyroxene granules. It

may be quite fresh but is more often altered to green serpentine. The optical character is invariably negative and 2V is always close to 87°, indicating a composition of $Fo_{80}Fa_{20}$ according to the diagram of Deer and Wager (1939, p. 21). There is a marked change in habit in the olivine diabase layer (Fig. 3), where the olivine occurs in two distinct generations. The larger crystals may exceed 2 mm. in diameter and may be sharply idiomorphic or partly resorbed. They are enclosed by all the major constituents and are not zoned. 2V varies between 87° and 82° indicating compositions between $Fo_{80}Fa_{20}$ and $Fo_{70}Fa_{30}$. The largest crystals are generally the poorest in fayalite.

Small crystals of olivine are very abundant and vary in diameter from 0.1 mm. to 0.02 mm. They may be rounded or, rarely, idiomorphic. The mineral tends to form loosely packed groups. It is distinctly richer in fayalite than the larger crystals and shows slight zoning with an increase in the proportion of fayalite toward the margin. Dr. Tomkeieff (1939, p. 229-251), who has kindly examined some of the slices, commented upon the zonal character of the smaller olivines, but the actual amount is difficult to determine owing to wedge effects near the margins. 2V varies from 82°-78° indicating $Fo_{70}Fa_{30}$-$Fo_{65}Fa_{35}$. This generation of small olivine crystals disappears northward and is less marked at Englewood Cliff and absent from the Alpine section.

The following are the optical properties of large and small olivine crystals from the middle of the olivine diabase layer half a mile south of George Washington Bridge. Throughout this paper the optical properties have been determined by monochromatic sodium light. Refractive indices were measured by the immersion method on crushed material or fragments isolated from the edges of the sections. The accuracy is ± 0.002. The optic axial angles were measured by universal stage methods with an accuracy of $\pm 2°$ and are calculated on the α index.

Large
2V (negative) = 84°
Refractive indices $\begin{cases} \alpha = 1.697 \\ \beta = 1.715 \\ \gamma = 1.731 \end{cases}$ $\gamma - \alpha = 0.034$

Small
2V (negative) = 78°
Refractive indices $\begin{cases} \alpha = 1.703 \\ \beta = 1.726 \\ \gamma = 1.740 \end{cases}$ $\gamma - \alpha = 0.037$

Both olivines are colorless.

PLAGIOCLASE

The proportion of plagioclase in the sill varies from 40-45 per cent at the lower contact to a minimum of about 32 in the olivine diabase layer above which it increases steadily to a maximum of 66 about 200 feet below the top of the sill. Thereafter it falls off again to 47 per cent near the upper contact.

The average length of the lath-shaped crystals, which are generally well formed, is given in the micrometric tables and in Figures 3 and 6; the maximum length is over 50 per cent greater. The ratio of length to breadth is normally between three and four but falls off to two in the olivine layer, where there is a sudden and very considerable increase in

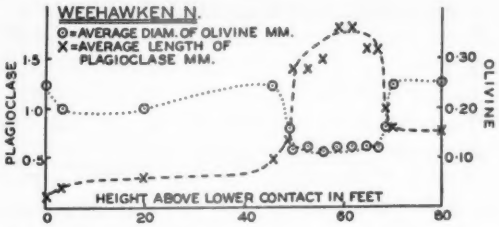


FIGURE 3.—*Average size of the olivine and plagioclase*
In the lower part of the section at Weehawken North

dimensions (Fig. 3). The increase becomes less marked at Englewood Cliff and negligible in the Alpine section.

There is nothing particularly noteworthy about the optical properties of the plaglioclase. Its composition, determined by extinction methods checked by refractive index measurements, varies according to the graph (Fig. 4). This curve represents the average composition of the central portions of the laths, but some zoning is always present. The zoning usually shows increasing soda content from center to margin and is very rarely oscillatory. The increase in soda is very slow in the central portions but becomes rapid near the margin which may consist of oligoclase. The most calcic variety measured was $Ab_{30}An_{70}$. Albite twinning is almost universal and is very often accompanied by Carlsbad twinning. Pericline twinning is rarer.

As has already been stated, the plagioclase crystallized after the olivine but is, on the whole, earlier than the pyroxene toward which it generally exhibits ophitic relationship, particularly in the case of the orthopyroxene.

In the most felsic portions of the sill there is a good deal of untwinned albite in the coarse schlieren, which is doubtless the product of hydrothermal activity. Granular albite containing practically no anorthite and showing fine twin lamellae is the chief constituent of the white residual injections.

The plagioclase is generally well preserved except in the pegmatitic schlieren where hydrothermal activity is pronounced. Elsewhere the composition gives rise to saussuritic products.

PYROXENE

General statement.—From every point of view the pyroxenes form the most interesting group of minerals in the sill. In variety of optical properties and chemical composition they are remarkable among rocks of tholeiitic magma type. Their mutual relationship, moreover, is decidedly

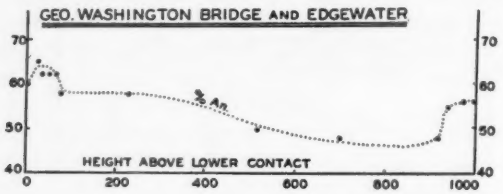


FIGURE 4.—*Anorthite content of plagioclase in the sill*
At George Washington Bridge and Edgewater

unusual and seems to be in conflict with recent work on similar rock types.

Several years before Lewis' detailed description it had been recognized that the Palisade diabase contained both ortho- and clinopyroxene (Merrill and others, 1902, p. 9; Lewis, 1908b, p. 156), however, disagreed with those determinations and stated categorically that all the pyroxene in the sill was augite and that what had previously been claimed as hypersthene was in reality only pleochroic augite. Later observations confirmed the presence of orthopyroxene in the sill.

The author records the presence of four distinct varieties of pyroxene in the diabase: pigeonite, hypersthene-augite, hypersthene, and clinohypersthene(?). There still remains a certain divergence of opinion as to the classification of the monoclinic pyroxenes. In this paper the term pigeonite is reserved for varieties with 2V less than 30°; all others are termed hypersthene-augite, since they are aluminous, and 2V never exceeds 50°.

Hypersthene-augite.—Though not the earliest pyroxene to crystallize, this is by far the most abundant. It is pale greenish-brown, the depth of tint tending to increase toward the top of the sill. The crystals vary from minute granules (0.01 mm. in diameter), in the groundmass of the contact modifications, to elongated prisms of 6—8 cm. in pegmatite schlieren. Idiomorphic phenocrysts occur sparsely in the basaltic contact phases and tend to form clusters. The mineral has the same optical properties in these as in the coarser diabase. Apart from the basaltic contact modifications it is generally subidiomorphic to subophitic and is definitely later than the calcic plagioclase. In the lower portions of the sill it is earlier than the

orthopyroxene, which frequently encloses it, but at higher levels where orthopyroxene is absent it may be preceded by pigeonite. Again, in the pyroxene-rich lower portions it tends to form stumpy prisms, while at higher levels, where the proportion is less, more elongated crystals occur. This is particularly the case in some of the pegmatitic schlieren which contain a large percentage of late micropegmatite. Elongated crystals may, however, be found at all levels and very often show slight curvature.

Simple twinning on (100) or (010) is common, and lamellar twinning less so. In the larger prisms in the felsic upper portion of the sill a well-marked basal parting and sahlite striation are developed and emphasized by inclusions of iron ore. All varieties of pyroxene tend to form polysomatic groups. The extinction angle of the hypersthene-augite (γ:c) is 44°, and 2V(+) is usually about 48°. The optical properties of a typical crystal from a point just above the olivine layer half a mile south of George Washington Bridge are as follows:

Refractive indices	$\alpha = 1.690$ $\beta = 1.699$ $\gamma = 1.714$	$\gamma - \alpha = 0.024$ $\gamma : c = 44°$	2V (positive) = 47° Color very pale green-brown

The composition according to Deer and Wager's (1938, p. 20-21) modification of Tomita's diagram is: $CaSiO_3$=30, $MgSiO_3$=44, $FeSiO_3$=26. Zoning is inconspicuous, but there is some variation of the optical properties in different crystals. 2V may fall to 35° in any part of the sill, and in the upper portions all varieties between hypersthene-augite and uniaxial pigeonite may be observed. The hypersthene-augite appears to be richer in the hedenbergite molecule toward the top, and refractive indices may be 0.005 higher than in the example quoted, but the axial angle is generally smaller. In the lower portions of the sill the mineral is usually fresh, but in the felsic upper portions hydrothermal alteration to amphibole or decomposition to carbonate are common. In the pegmatite schlieren the hypersthene-augite often shows graphic intergrowth with plagioclase which forms the host. This intergrowth is generally associated with hydrothermal activity, and the pyroxene is often altered to brown amphibole. In transverse sections the pyroxene rods exhibit typical cuneiform habit (Pl. 2, fig. 1).

Pigeonite.—In the upper half of the sill much of the pyroxene is characterized by very low axial angles; 2V in many cases approaches zero. This variety is undoubtedly pigeonite. The optical properties are otherwise quite similar to those of the hypersthene-augite, but the double refraction and occasionally the extinction angle are slightly lower. Pigeonite may occur as individual crystals with the same habit as the hypersthene-augite and sometimes is enclosed in it, or the two varieties may be

associated in the same crystal, in which case the boundary (as shown by the difference of extinction and double refraction) is often quite sharp, and the pigeonite invariably is in the center. The two varieties have the same refractive index, there is no Becke effect, and cleavage cracks pass from one into the other. One example near the electric railroad track at Fort Lee showed a sharply defined uniaxial core surrounded by a variety with 2V=48°. It appears, however, that the series pigeonite-hypersthene-augite is continuous, for all intermediate varieties may be found in the upper parts of the sill. The optical properties of pigeonite from the above locality at Fort Lee—*i. e.*, just above the center of the sill—are:

Refractive indices $\begin{cases} \alpha = 1.689 & 2V\,(+) = 0° \\ \beta = 1.689 & \gamma - \alpha = 0.021 \text{ Pale green-brown} \\ \gamma = 1.710 & \gamma : c = 44° \end{cases}$

The composition according to Deer and Wager's modification of Tomita's diagram is: $CaSiO_3$=13, $MgSiO_3$=53, $FeSiO_3$=34.

Hypersthene.—Orthopyroxene occurs in both the upper and lower chilled phases but otherwise appears to be confined to the lowest third of the intrusion. In the basaltic contact modification it is found as minute granules surrounding olivine, and, as Lewis (1908b, p. 156) has stated, is obviously a reaction product. In the coarser varieties, however, it may be abundant. It is readily distinguished from the clinopyroxene by its straight extinction, lower double refraction, and characteristic pink-green pleochroism. In habit it is somewhat unusual for, although subidiomorphic prisms are found, it tends rather to form large irregular plates up to 3 mm. across. These are usually untwinned and bear an ophitic or, as Lewis terms it, a "micro-poikilitic," relationship toward the plagioclase laths. It crystallized late and frequently encloses small prisms of earlier hypersthene-augite. Strain is indicated in places by undulatory extinction without change of the axial angle. The optical character is invariably negative, and 2V=69° to 71°. In some crystals there is a thin margin of higher double refraction and lower axial angle, which is clearly richer in ferrosilite. A crystal from the center of the olivine diabase layer half a mile south of George Washington Bridge has the following optical properties:

Refractive indices $\begin{cases} \alpha = 1.690 & 2V\,(-) = 72° \\ \beta = 1.698 & \gamma - \alpha = 0.013 \\ \gamma = 1.703 & \gamma : c = 0° \end{cases}$ Pleochroism $\begin{cases} \alpha = \text{pale pink} \\ \beta = \text{pale yellowish-pink} \\ \gamma = \text{pale green} \end{cases}$

These data indicate a hypersthene with 35 per cent ferrosilite, according to the curves of Henry (1935, p. 223). There is little variety of composition in the central portions of the crystals, but in the margins the proportion of ferrosilite may exceed 40 per cent (Scholtz, 1936, p. 104). Alteration to green bastite or serpentine is usually slight.

Clino-hypersthene (?).—Many pyroxene plates in the lowest third of the intrusion have the pleochroism and low double refraction of hypersthene but show extinction angles up to 30°. Closer examination of these crystals proves that they are crowded with rodlike inclusions of clinopyroxene with extinction angles ranging up to 44°. These, like the lamellae in orthopyroxene described by Hess and Phillips (1938, p. 450-456), doubtless account for the oblique extinction. Their rodlike habit is, however, quite different from that of the lamellae in orthopyroxenes of the Bushveld type. Lewis (1908a, p. 110) described and figured them as interpenetration twinning and commented upon their resemblance to the graphic intergrowth of quartz and orthoclase. This resemblance is indeed striking, and the author is tempted to ascribe the origin of the rodlike inclusions to graphic intergrowth of ortho- and clinopyroxene as a result of unmixing. Henry (1938, p. 23-29), who has recently reviewed the data relating to clinohypersthenes and has kindly examined some of the author's sections, ascribes them to disorientation through strain. They are too small to yield interference figures but since they are associated with late orthopyroxene they probably consist of hypersthene-augite rather than of pigeonite. They are elongated parallel to the c axis.

This curious intergrowth or disorientation effect may be confined to the center or to one side of crystals which in their other portions consist of normal hypersthene. It is found at all levels where hypersthene is recorded.

Relative abundance and distribution of ortho- and clinopyroxene.—Where both ortho- and clinopyroxenes occur either may predominate, though the proportion is usually fairly equal. In the coarser upper portions hypersthene-augite, normally of somewhat low axial angle, is generally more abundant than pigeonite.

The total pyroxene percentage by weight reaches a maximum of over 60 near the olivine diabase layer—usually a little above it—and falls off progressively to a minimum of 30 about 200 feet below the upper contact. The chilled phases contain slightly under 50 per cent.

Hypersthene and pigeonite seem to be incompatible in the unchilled diabase. The former is not found above 400 feet from the lower contact, and the latter rarely below this level. The complementary nature of plagioclase and pyroxene in the sill is well shown in Figures 5 and 6.

Composition of pyroxene.—The composition of the pyroxenes has been deduced from their optical properties alone. Owing to their intimate association and lack of variation in density, mechanical separation proved impracticable.

An analysis of pyroxene, separated by Phillips (1899, p. 281) from the upper parts of the Rocky Hill sill, which may be a continuation of the Palisade diabase, shows a great excess of ferrous iron and magnesia over lime ($FeO = 18.34$, $MgO = 12.89$, $CaO = 11.40$), probably indicating the presence of pigeonite along with augite.

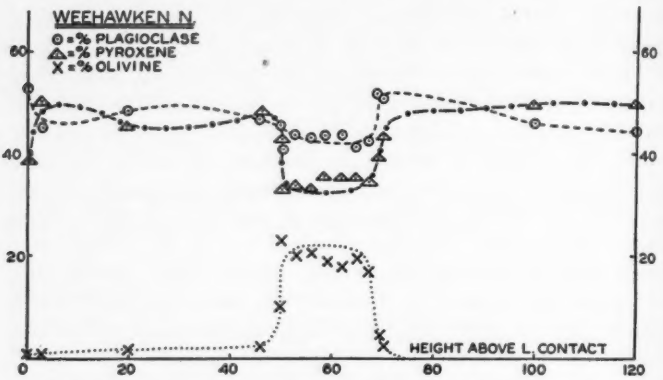


FIGURE 5.—*Mineral composition of the lower part of the section*
At Weehawken North

MINOR CONSTITUENTS

Iron ore.—Titaniferous iron ore is present throughout the sill and in the coarsest upper portions is sufficiently abundant to be a major constituent. In the rapidly cooled contact modifications small well-formed octahedra of early crystallization are prominent; in the coarser central portions the mineral is more skeletal and is of somewhat later crystallization. In the pegmatite schlieren below the upper chilled phase, it occurs as irregular skeletal masses up to 4 or 5 mm. in length and of still later formation. In this last case the mineral is undoubtedly ilmenite and may occasionally show alteration to leucoxene. Throughout the lower two-thirds of the sill, including the olivine diabase layer, iron ore is distinctly scarce and it may be completely absent from a field 3 mm. in diameter.

Scaly pyrite is found in the contact modifications near joints.

Biotite.—Reddish or chocolate-brown biotite is sparse throughout the sill but may show local concentrations owing to hydrothermal reactions. It occurs as small irregular flakes, generally moulded on pyroxene or iron ore, and probably represents a late reaction product of these minerals.

The proportion increases near joint fissures, some of which are lined with black mica. In the upper chilled phase the mineral becomes prominent and assumes a well-marked poikilophitic habit, forming irregular plates up to 2 mm. across. Here, too, it seems to be the result of migratory

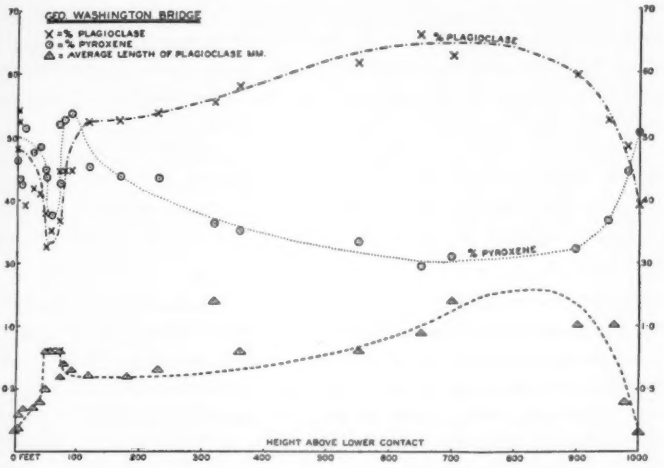


FIGURE 6.—*Amount of plagioclase and pyroxene and size of plagioclase* In section at George Washington Bridge

hydrothermal solutions, rich in potash, reacting with the mafic constituents. The following are the optical properties of biotite from the upper chilled phase about 10 feet below the upper contact at the intersection of Jones Road and Highway 4:

Refractive indices $\begin{cases} \alpha = 1.591 \\ \beta = 1.641 \\ \gamma = 1.641 \end{cases}$ $2V\,(-) = 0°$ $\gamma - \alpha = 0.040$ Pleochroism $\begin{cases} \alpha = \text{pale yellow-brown} \\ \beta = \text{dark red-brown} \\ \gamma = \text{deep chocolate-brown} \end{cases}$

Micropegmatite.—Graphic intergrowth of quartz and alkalic feldspar may be found at all levels in the sill, except in the basaltic contact modifications and in the olivine diabase layer or its immediate vicinity. The proportion of micropegmatite in the normal diabase is generally small but shows a progressive increase upward until the chilled phase is reached; the maximum percentage by weight is about 6. In the pegmatite schlieren, however, as much as 25 per cent was recorded, while the figures quoted by Lewis (1908a, p. 123) indicate even larger proportions.

The size of the rodlike quartz inclusions is very variable but increases with the grain of the rock, so that in the coarsest modifications it is

possible to determine the nature of the alkalic feldspar which occupies the interspaces. It appears to be a micro- or kryptoperthite with 2V(—) about 40° (*cf.* Scholtz, 1936, p. 117). The mean refractive index is 1.528.

The proportion of quartz in the micropegmatite varies from 40 to 50 per cent, and the intergrowth is invariably interstitial, representing the last constituents to crystallize. In some of the pegmatite schlieren it has actually worked along the cleavage cracks of the basic plagioclase crystals and has replaced their centers in a manner reminiscent of analcite in teschenites. In this case it is doubtless hydrothermal.

In the rapidly cooled chilled phases either quartz or alkalic feldspar may exist independently as an interstitial mineral, and elsewhere quartz alone, but micrographic intergrowths are far more common. Some of the alkalic feldspar is fresh, but most of it shows turbid alteration products.

Amphibole.—Amphibole is confined to the coarse and relatively siliceous modifications of the pegmatite zone and seems to be the result of late magmatic or hydrothermal reaction with pyroxene. It is usually associated with patches of micropegmatite and often forms a marginal growth on crystals of clinopyroxene. Its optical properties are exceedingly variable. The commonest type is a greenish-brown hornblende which is strongly pleochroic in shades of these colors and has low extinction angles (γ: c = 12° — 16°). Reddish-brown basaltic hornblende rich in Fe_2O_3 and blue alkaline varieties apparently allied to arfvedsonite are also common. The latter show the following optical properties which agree well with arfvedsonite:

$$\text{Pleochroism} \begin{cases} \alpha = \text{green-yellow} & \gamma : c = 25° \\ \beta = \text{olive-green} & 2V\,(-)\ \text{approx.} = 40° \\ \gamma = \text{green-blue} \end{cases}$$

All varieties are most irregular in habit and optical properties. Groups of small crystals are common, and in some cases, as mentioned above, micrographic clinopyroxene has been entirely replaced by brown hornblende.

Accessories.—Apatite is ubiquitous but is most conspicuous in the coarser rocks where the crystals are relatively large. It occurs as slender colorless needles of early crystallization.

An interesting accessory, occurring only in the coarsest pegmatite schlieren, is pectolite which forms spherical aggregates of radiating fibers with low extinction and positive elongation; $\gamma = 1.632$, $\alpha = 1.601$. The mineral has optical properties very similar to those of pectolite recorded by O. Pácak (1926) in a picrite from Moravia.

MICROMETRIC DATA

ACCURACY

Since many of the conclusions drawn in this paper are based on micrometric analyses of thin sections, it is desirable to state the precautions taken to ensure accuracy.

Apart from the basaltic contact modifications which are too fine and the pegmatite schlieren which are too coarse, the grain size of the Palisade diabase is very favorable to accurate micrometric measurement. The texture is, on the whole, fairly even, and none of the grains is very large. Most of the author's sections have an area of at least 500 sq. mm. and contain no holes. This means that the area of the section is at least 200 times greater than that of the largest grain which it contains. Examination of duplicate sections showed that the sampling of the sill at the various levels was representative. Most of the measurements were made with a Dollar integrating stage which was found very suitable for the purpose, since it permits of long traverses, and the proportions of six minerals can be estimated simultaneously. A careful examination of the entire section was always made before measuring it micrometrically, and doubtful minerals were identified. This reduced to a minimum problems of identification during the actual traverses when convergent light was not available. It may thus be affirmed that in nonpegmatitic rocks at distances of more than 10 feet from the contacts the results have an accuracy approaching 1 per cent and certainly not exceeding 2 per cent. (*See* Larsen and Miller, 1935, p. 138-146). This was confirmed by duplicate determinations in several cases. The results, except in the case of abnormally coarse or fine rocks, are given to the first decimal place, but in the percentage of the major constituents the decimal figure must be regarded only as a control of the integer. Several of the finest-grained rocks from the chilled phases and residual injections were measured by means of a six-spindle Leitz-Scheumann stage which is more suitable for rocks of small grain size. In such cases a probable accuracy of only 2 or 3 per cent was obtained, and the estimations are marked by an asterisk. In all the measured sections except those of pegmatite schlieren the length of traverse was over 200 times greater than that of the longest continuous intercept made by any mineral grain.

Micrometric analyses of the pegmatite schlieren are given to the nearest per cent, but the accuracy attained is probably not more than 5 per cent. Not only is the grain size large but the texture is uneven and patchy, and a good deal of alteration has taken place. This also applies to the white residual injections, which are of fine grain and carry considerable altered and unidentifiable alkalic feldspar.

DATA

In the micrometric data given in tables, the position of each slice in the sill is given in feet above the lower contact. Where the contact is visible and there is a well-exposed cliff section above, estimates up to the crest of the cliff may be made fairly accurately with the use of an aneroid barometer. Up to 150 feet above the lower contact the error probably does not exceed 10 feet. Back from the cliff edge, however, it is more difficult to determine the height of a specimen above the lower contact. In practice the location was easily fixed in the field on Hagstrom's street map of Bergen County, for most of the ground not yet built upon is marked out in streets. The locality was then transferred to the 3 inch = 1 mile enlargement of the United States Geological Survey 1:62,500 topographic sheet. The dip of the sill was known, and the height of the specimen above the lower contact could be calculated. This method is dependent on the accuracy of the contouring (20-foot intervals) and on the lower contact being nontransgressive. Considerable errors up to 50 feet or more may be introduced in localities in the upper portions of the sill. Toward the top, however, where the specimens may be referred to a visible upper contact, the error becomes at least more consistent. The thickness of the sill along its outcrop is variable, but most estimates published place it at about 1000 feet at Fort Lee (875 feet proved by a bore hole sunk in diabase); it tends to thin out toward Jersey City. The figures quoted in the tables were arrived at by consideration of all the available evidence and by calculations based on the dip and width of outcrop, due allowance being made for faulting. The probable error is less than 100 feet.

Series of specimens were collected from ten of the best exposed sections through the sill between Alpine and Hoboken. Four of the sections (4a, 4b, 5a, 5b) are less complete, and for the sake of brevity their measurements are omitted. They yielded results which were entirely in accord with the other sections, however. Numerous isolated specimens were collected from localities between the numbered sections. All the specimens from the serial sections were sliced and measured micrometrically. The percentage mineral composition by volume for six sections is given in the table.

In Table 2 are given micrometric measurements of the pegmatitic schlieren and white veins.

CHEMICAL DATA

Lewis' (1908a, p. 121) account of the Palisade diabase is supplemented by ten excellent chemical analyses by R. B. Gage. The exact positions in the section are listed only for the basaltic contact modifications and

TABLE 1.—*Micrometric measure*

Height in feet above lower contact	Olivine	Pyroxene	Plagioclase	Iron ore	Biotite	Quartz and micropegmatite	Color ratio	Average diameter in mm. of olivine	Average length in mm. of plagioclase
				(1) Alpine					
175	.....	41.6	53.7	1.5	1.1	2.1	44.2	.....	0.6
120	.....	42.4	52.4	2.7	0.6	1.9	45.7	.....	0.6
67	2.4	45.7	49.2	1.3	0.8	0.6	50.2	0.18	0.6
60	11.2	46.2	40.6	1.4	0.6	.....	59.4	0.19	0.6
55	12.2	44.3	40.8	1.6	1.1	.....	59.2	0.18	0.6
42	4.8	48.2	44.6	2.3	0.1	.....	55.4	0.20	0.5
30	2.9	43.9	50.2	2.0	1.0	.....	49.8	0.20	0.25
10	2.3	44.1	51.8	1.5	0.3	.....	48.2	0.20	0.25
				(2) Englewood					
990	0.3	47.3	47.2	2.0	1.8	1.4	51.4	0.20	0.25
960	1.4	44.7	49.8	2.2	0.5	1.4	48.8	0.20	0.35
940	.....	38.3	55.9	3.6	0.4	1.8	42.3	.....	0.80
900	.....	36.0	55.2	3.4	0.6	4.8	40.0	.....	0.80
500	.....	28.4	63.7	4.0	0.1	3.8	32.5	.....	0.70
400	.....	30.0	63.5	2.1	0.2	4.2	32.3	.....	0.70
225	.....	36.9	56.4	4.0	0.2	2.5	41.1	.....	0.65
180	.....	39.8	55.5	1.9	0.4	2.4	42.1	.....	0.65
140	.....	44.6	51.5	2.3	0.2	1.4	47.1	0	0.60
100	0.1	49.9	46.4	2.6	0.4	0.6	53.1	0.20	0.60
70	19.8	41.5	34.8	2.0	1.9	.....	65.2	0.16	0.60
64	17.8	42.6	36.2	1.9	1.5	.....	63.8	0.20	0.60
61	16.0	42.5	37.4	1.8	2.3	.....	62.6	0.15	0.70
55	9.1	46.6	42.7	1.7	0.9	.....	57.3	0.25	0.40
45	2.4	47.6	48.1	1.4	0.2	0.3	51.6	0.25	0.35
30	0.4	46.5	49.2	2.8	0.3	0.8	50.0	0.25	0.35
6	1.4	46.4	49.4	2.3	0.2	0.3	50.3	0.20	0.30
1	2.2	47.2	46.0	4.6	.....	.....	54.0	0.20	0.20
				(3) George Washington Bridge					
997	0.7	44.4	49.4	2.9	2.4	0.2	50.4	0.20	0.18
990	0.7	44.4	48.2	3.0	2.3	1.4	50.4	0.25	0.40
950	.....	36.8	52.4	3.7	0.2	6.9	40.7	.....	1.00
900	.....	32.0	59.9	1.6	0.3	6.2	33.9	.....	1.00
700	.....	30.7	62.6	1.5	0.6	4.6	32.8	.....	1.20
650	.....	29.4	65.8	0.9	0.1	3.8	30.4	.....	0.90
550	.....	33.0	61.3	1.7	0.4	3.6	35.1	.....	0.80
320	.....	36.0	55.1	1.0	3.8	4.1	40.8	.....	1.20
220	.....	43.1	53.5	1.0	0.4	2.0	44.5	.....	0.65
170	.....	43.7	52.2	1.0	1.7	1.4	46.4	.....	0.60
120	.....	45.0	52.2	1.3	0.7	0.8	47.0	.....	0.60
90	0.2	53.4	44.2	1.3	0.3	0.6	55.2	0.20	0.65
80	.....	52.7	44.4	1.3	0.4	1.2	54.4	.....	0.70
73	1.0	51.8	44.5	1.9	0.6	0.2	55.3	0.20	0.60
71	16.5	42.5	36.6	1.6	0.8	.....	61.4	0.14	0.80
60	22.2	37.4	34.9	2.5	3.0	.....	65.1	0.12	0.80
50	21.2	45.7	32.6	1.2	1.5	.....	67.4	0.14	0.80
49	14.5	44.6	37.8	2.2	0.9	.....	62.2	0.20	0.50
40	8.1	47.1	41.9	2.1	0.7	0.1	58.0	0.20	0.40
30	8.8	48.5	41.0	1.2	0.5	.....	59.0	0.20	0.35
15	7.1	51.2	39.2	1.8	0.6	0.1	60.7	0.20	0.35
6	2.0	42.6	52.2	1.6	0.4	1.2	46.6	0.20	0.30
3	1.3	43.0	54.2	1.3	0.1	0.1	45.7	0.20	0.20
1	1.9	46.3	48.0	2.6	1.0	0.2	51.8	0.20	0.18

(1) *Alpine.* Descent to Yonkers Ferry from Palisades Boulevard to the lowest exposed diabase. Sill 1,000 feet thick. Olivine layer 50–65 feet above contact.
(2) *Englewood Cliff.* Descent to Dyckman Street Ferry from the lower contact to the crest of the cliff; thence by Palisade Avenue, Woodland Street, Linden Avenue to the upper contact by Flatrock Brook at the intersection of Highway 4 and Jones Road. Sill 1000 feet thick. Olivine layer 60–80 feet above contact.
(3) *George Washington Bridge.* From lower contact on the Hendrik Hudson Drive, 600 yards south of George Washington Bridge, up the Bluff behind; thence by Central Avenue to Taylorville Hollow and along Highway 4 to West View Avenue. Sill 1000 feet thick. Olivine layer 50–70 feet above contact.
(4) *Edgewater Cliff.* From lowest exposed diabase on descent to 125th Street Ferry to crest of cliff; thence by Bluff Road to Palisade Avenue and Euclid Street. Sill 1000 feet thick. Olivine layer 60–75 feet above contact.

ıetric measurements

Height in feet above lower contact	Olivine	Pyroxene	Plagioclase	Iron ore	Biotite	Quartz and micropegmatite	Color ratio	Average diameter in mm. of olivine	Average length in mm. of plagioclase
				(4) Edgewater					
550	.....	30.3	60.8	5.1	0.4	3.4	35.8	.....	1.00
350	.....	35.0	57.9	3.3	0.4	3.4	38.7	.....	0.85
260	.....	39.3	55.5	1.5	0.9	2.8	41.7	.....	0.75
200	.....	41.1	54.9	1.5	0.7	1.8	43.3	.....	0.70
135	.....	41.7	52.9	1.8	0.3	3.3	43.8	.....	0.80
115	.....	42.2	55.3	1.9	0.2	0.4	44.1	.....	0.80
95	.....	45.0	52.6	1.9	0.2	0.3	47.1	.....	0.80
80	0.5	50.2	46.1	1.6	0.3	0.3	53.6	0.25	0.60
77	7.2	49.9	39.5	3.0	0.4	.....	60.5	0.25	0.60
74	16.0	42.5	37.5	2.0	2.0	.....	62.5	0.15	1.20
72	25.5	38.4	34.0	0.6	1.5	.....	66.0	0.12	1.60
70	23.2	41.0	33.4	1.9	1.5	.....	66.6	0.12	1.80
68	21.0	42.6	31.6	2.2	2.6	.....	68.4	0.12	1.60
63	24.0	39.1	34.1	1.1	1.7	.....	65.9	0.15	1.40
61	14.5	48.4	33.1	1.6	2.4	.....	66.9	0.18	0.70
60	7.9	45.6	43.6	2.0	0.9	.....	56.4	0.20	0.50
57	5.4	49.3	42.7	2.2	0.4	.....	57.3	0.20	0.35
35	4.5	50.5	42.7	2.1	0.2	.....	57.3	0.20	0.35
20	5.0	49.1	44.2	1.3	0.4	.....	55.8	0.18	0.30
4	1.8	45.5	50.0	2.7	.....	.....	50.0	0.20	0.20
1	2.9	50.9	42.4	3.0	0.6	0.2	57.4	0.25	0.12
				(5) Weehawken N.					
795	0.6	45.4	50.8	3.2	.....	.....	49.2	0.30	0.15
785	0.6	44.7	51.1	2.4	.....	1.2	47.7	0.30	0.50
750	.....	40.7	54.9	2.3	0.3	1.8	43.3	.....	0.60
700	.....	35.0	58.6	1.7	0.1	4.6	36.8	.....	0.80
550	.....	33.5	57.3	3.0	0.3	5.9	36.8	.....	0.95
120	.....	44.4	49.7	2.9	0.9	2.1	48.2	.....	0.80
100	.....	45.9	49.0	2.4	1.1	1.6	49.4	.....	0.80
70	1.5	50.7	43.5	2.5	0.3	1.5	53.5	0.25	0.80
69	4.3	51.3	39.5	1.5	3.4	.....	56.2	0.16	1.00
67	16.7	42.4	34.6	2.6	3.7	.....	65.4	0.12	1.60
65	19.7	41.4	35.6	2.1	1.2	.....	64.4	0.12	1.60
62	17.8	43.5	35.2	2.3	1.2	.....	64.8	0.12	1.80
59	19.0	43.2	35.4	1.3	1.1	.....	64.6	0.12	1.80
56	20.4	42.8	33.0	1.8	2.0	.....	67.0	0.13	1.50
53	20.0	43.8	33.6	1.3	1.5	.....	66.4	0.12	1.40
50	23.0	41.2	32.9	1.6	1.3	.....	67.1	0.12	1.40
50	10.0	45.1	42.6	1.8	0.5	.....	57.4	0.16	0.70
46	2.4	47.3	47.9	1.7	0.5	0.2	51.9	0.24	0.50
20	2.0	48.2	45.4	3.9	0.1	0.4	54.2	0.20	0.30
3	1.0	44.9	50.0	3.6	0.3	0.2	49.8	0.20	0.20
0	1.2	52.5	39.2	6.3	0.8	.....	60.8	0.23	0.10
				(6) Hoboken					
120	.....	45.3	47.6	1.6	1.2	4.3	48.1	.....	1.00
93	.....	47.0	46.4	1.9	0.8	3.9	49.7	.....	0.85
80	8.4	49.3	39.6	2.2	0.5	.....	60.4	0.15	0.80
79	14.1	43.3	39.3	1.7	1.6	.....	61.7	0.12	1.20
78	22.3	41.8	33.2	1.2	1.5	.....	66.8	0.12	1.40
70	23.4	40.5	32.8	1.2	2.1	.....	67.2	0.12	1.50
60	23.5	40.4	32.9	1.2	2.0	.....	67.1	0.13	1.40
59	12.5	43.8	41.3	1.0	1.4	.....	58.7	0.16	0.80
52	6.7	49.1	40.6	2.6	1.0	.....	59.4	0.20	0.50
48	5.8	51.6	40.0	1.8	0.8	.....	60.0	0.20	0.50
30	1.4	54.6	39.3	2.3	1.0	1.4	53.3	0.30	0.50

(4a) *Edgewater Electric Railroad.* From lower contact on electric railroad descent to 125th Street Ferry to intersection with Bluff Road.

(4b) *Edgewater-Fairview Tunnel.* From lower contact on River Road at entrance to Edgewater-Fairview tunnel up slope to quarries on Gorge Road.

(5) *Weehawken North.* From lower contact on descent from Eastern Boulevard to River Road up to Eastern Boulevard; then north along Boulevard to Bull's Avenue and along 6th Street to Kamena Street; thence by Dallytown Road and finally over ridge to upper contact above intersection of Prospect and Fairview Avenues. Sill 800 feet thick. Olivine layer 50–68 feet above contact.

(5a) *Weehawken.* From lowest exposed diabase on Pershing Road to crest of cliff.

(5b) *King's Point.* From lower contact above railroad tracks at King's Point to top of bluff.

(6) *Hoboken.* From lowest exposed diabase south of Paterson Plank Road to crest of cliff. Sill 700 feet thick. Olivine layer 60–78 feet above contact.

the olivine layer, however. Moreover, the specimens analyzed were from points as far apart as Jersey City and Englewood Cliff. It seemed, therefore, advisable in the present investigation to include a series of reliable analyses of specimens from the most complete section avail-

TABLE 2.—*Micrometric measurements*

	Pegmatite schlieren			White veins		
	1	2	3	4	5	6
Quartz	5	6	4	12	2	12
Alkalic feldspar	4	4	3	70	78	80
Calcic plagioclase	50	58	64	.....	.....	.....
Pyroxene	30	21	20	15	18	5
Amphibole	3	4	3	.....	.....	.....
Iron ore	6	5	5	1	1	1
Biotite	2	2	1	.....	.....	.....
Sphene	.....	.....	.....	1	1	2
Calcite	.....	.....	.....	1	.....	.....
Average length in mm. of feldspar	2.5	2.5	3.0	0.40	0.35	0.30

1. George Washington Bridge, 850 feet.
2. Weehawken North, 700 feet.
3. Englewood, 900 feet.
4. George Washington Bridge, 20 feet.
5. Weehawken North, 20 feet.
6. Weehawken North, 720 feet.

able, the exact position in the section being noted in each case. Fifteen such analyses were made for the author by Mr. F. A. Gonyer of Harvard University and by Mr. F. Herdsman; they were found to confirm the micrometric data in every particular. The section selected was No. 3, George Washington Bridge, since it extends from a well-exposed lower contact to within 3 feet or so of the upper contact, contains no large gaps, and shows clearly the relationship between the olivine diabase layer and the normal diabase. The analyses are quoted in Table 3, together with the norms, specific gravities, and modes (calculated from the micrometric data). In Table 4 all R. B. Gage's analyses are quoted, with the exception of those where the exact locality is doubtful, and for comparative purposes some others of similar magma type are given. The close resemblance between analyses 3, 8, and 10 in Table 4 is noteworthy.

PETROLOGY

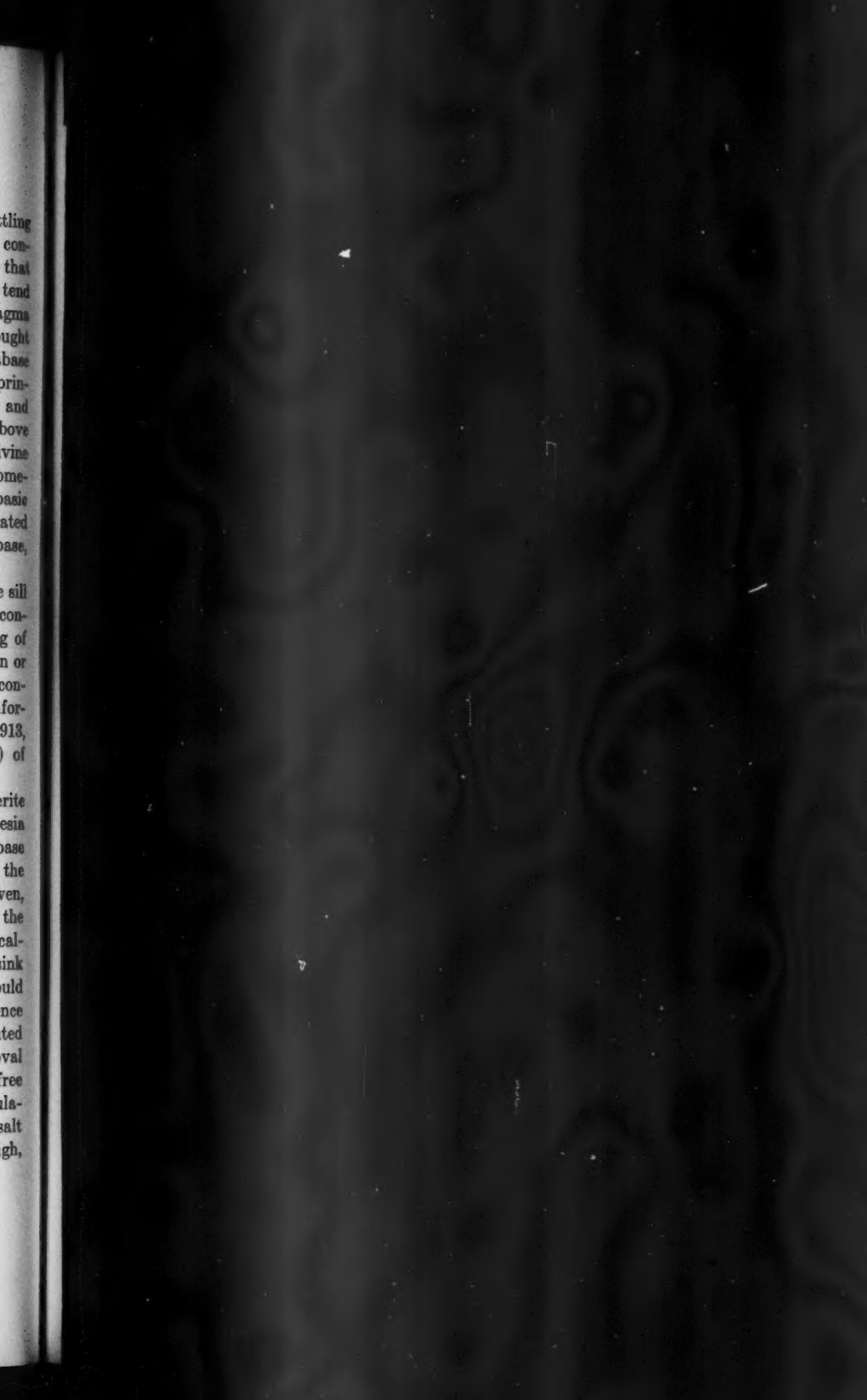
HISTORY OF INVESTIGATION

Lewis (1908a; 1908b) was the first to discuss in detail the differentiation of the Palisade diabase, and his account has the great advantage of being based on considerable chemical and micrometric data. He

attributed the main differentiation of the sill to the gravitational settling of iron ore and olivine crystals, which accumulated on the lower contact phase to form the olivine diabase ledge. He also realized that the early feldspar crystals would tend to rise and the augite would tend to sink, but he considered that the increasing viscosity of the magma would prevent this being brought about to any great extent. He thought the basaltic contacts were slightly more basic than the average diabase magma and that they originated through the operation of Soret's principle. He regarded the sill as having been injected in a single stage and noted, in this connection, the uniform height of the olivine layer above the lower contact along the outcrop. The contacts between the olivine diabase layer and the diabase above and below were reported as sometimes sharp and sometimes gradational; he noted further that the basic layer, which was invariably coarse-grained, was occasionally penetrated by what appeared to be an apophysis of the adjacent fine-grained diabase, but he advanced no explanation of these relations.

Since the publication of this detailed and well-reasoned account, the sill has been cited in numerous papers and textbooks as one of the most convincing examples of gravitational differentiation through the settling of early olivine crystals. Only such references as offer critical discussion or throw fresh light upon the subject will be considered here. No such contributions appeared for 7 years, but during this period important information as to the intrusion temperature (Sosman and Merwin, 1913, p. 389-395) and density (Day, Sosman, and Hostetter, 1914, p. 34) of the Palisade magma became available.

Bowen (1915, p. 168-191) showed empirically that crystals of forsterite and pyroxene sink in liquids with the composition of lime-magnesia pyroxene plus a little silica. He considered that the Palisade diabase provided an excellent natural example of this process and that the olivine layer accumulated in the manner suggested by Lewis. Bowen, however, thought it unnecessary to invoke the Soret principle for the formation of the basaltic contacts. Applying Stokes' formula he calculated that the olivine crystals with a diameter of 0.7 mm. would sink at the rate of about one meter per hour, and that 200 to 300 hours would suffice for the accumulation of the olivine layer, assuming the absence of convection currents which he regarded as unimportant. He attributed some, at least, of the quartz in the upper portions of the sill to the removal of olivine crystals, which would otherwise have reacted with the free silica to form pyroxene. The viscosity of the magma in Bowen's calculation was assumed to be eight times greater than that of Hawaiian basalt (Becker), but he pointed out that this value might well be too high, owing to the presence of water and other volatile constituents.



Ge

Height in feet above lower contact	
SiO_2	
TiO_2	
Al_2O_3	
Fe_2O_3	
FeO	
MnO	
MgO	
CaO	
Na_2O	
K_2O	
H_2O+	
H_2O-	
P_2O_5	
S	
CO_2	
Specific gravity	

Quarts		
Orthoclase		
Albite		
Anorthite		
Diopside	$CaSiO_3$	
	$MgSiO_3$	
	$FeSiO_3$	
Hypersthene	$MgSiO_3$	
	$FeSiO_3$	
Olivine	Mg_2SiO_4	
	Fe_2SiO_4	
Acmite		
Magnetite		
Ilmenite		
Apatite		
Calcite		
Water		

Quarts	
Orthoclase	
Plagioclase	
Pyroxene	
Olivine	
Iron Ore	
Biotite	
Amphibole	
Calcite	
Sphene	

* Analysis by F. Herdan

TABLE 3.—*New analyses*

George Washington Bridge section. Thickness of sill = 1000 feet. Olivine layer 50–70 feet above lower contact.
(Unless noted otherwise, analyses are by F. A. Gonyer.)

t in feet above wer contact	1	30	40	60	73*	80	220	500*	700	900	950	995*	Peg-matite* 600	White vein 20	Arkos —3
	51.82	49.12	50.81	48.28	50.42	51.72	51.82	52.32	51.46	51.48	52.16	52.02	52.50	72.52	72.6
	1.34	0.98	0.74	0.82	0.87	1.01	0.93	0.97	2.38	1.34	1.28	1.06	1.36	0.42	0.5
	14.76	11.82	9.42	9.36	14.01	14.36	15.80	16.54	16.24	15.22	15.96	15.56	18.28	11.84	13.8
	1.22	2.19	1.68	2.14	1.03	1.82	1.98	1.58	3.04	2.62	1.92	0.38	1.97	1.23	0.3
	9.24	9.62	9.10	11.54	8.84	8.22	7.70	8.66	9.58	9.67	8.51	9.54	7.61	0.87	0.4
	.12	0.14	0.11	0.12	0.03	0.07	0.09	0.12	0.05	0.09	0.06	0.03	0.09	0.06	0.0
	7.32	13.44	15.01	17.48	10.66	8.92	7.21	5.43	3.47	5.39	5.51	7.46	2.58	1.48	1.2
	10.02	9.14	10.65	7.00	10.04	9.10	10.40	9.68	8.38	8.96	10.01	9.12	9.34	2.54	1.0
	2.06	1.65	1.18	1.59	1.59	1.93	2.40	2.32	3.01	2.78	2.36	2.77	3.78	7.12	5.5
	0.82	0.55	0.40	0.41	0.60	1.08	0.87	1.03	0.87	1.21	0.77	0.68	1.03	0.36	3.5
	0.84	0.99	0.92	0.99	1.22	1.15	0.73	0.84	0.87	1.11	0.73	0.98	1.06	0.56	0.3
	0.14	0.05	0.06	0.06	0.48	0.26	0.27	0.40	0.26	0.26	0.33	0.20	0.50	0.26	0.1
	0.12	0.14	0.16	0.11	0.04	0.12	0.11	0.06	0.12	0.09	0.13	0.27	0.06	0.14	0.0
	none	none	none	none	none	none	none	none	none	none	none	none	none	0.07	0.0
	.....	.....	.....	.....	trace	.....	.....	0.16	.....	.....	.....	trace	none	0.44	.....
	99.82	99.83	100.16	99.90	99.83	99.76	100.31	100.11	99.73	100.22	99.73	100.07	100.16	99.94	99.8
ity.....	3.00	3.12	3.12	3.15	3.14	3.14	3.00	2.97	2.91	2.99	3.00	3.01	2.91	2.67	2.4

Norms

	2.5	.....	.....	.....	.....	2.0	1.4	4.0	5.4	0.9	4.6	.....	2.2	25.7	23.3
	5.0	2.8	2.2	2.2	3.3	6.7	5.6	6.1	5.6	8.3	4.5	3.9	6.1	2.2	21.1
	17.8	13.6	10.0	13.6	13.6	16.2	20.4	19.4	25.1	23.6	19.9	23.6	32.0	58.6	47.2
	28.4	23.6	19.2	17.2	29.2	27.0	29.5	31.7	28.1	24.7	30.9	28.1	29.8	.....	2.2
aSiO$_3$	8.6	8.7	13.7	6.7	8.2	7.3	8.9	6.0	5.3	7.9	7.5	6.5	6.5	3.8	0.8
gSiO$_3$	4.7	5.7	3.4	4.6	5.0	4.3	5.3	3.0	2.3	3.9	3.9	3.5	2.6	0.4	0.7
eSiO$_3$	3.6	2.4	9.2	1.9	2.8	2.6	3.2	2.9	3.0	3.8	3.4	2.8	4.0	3.0	.....
{ MgSiO$_3$	13.6	19.4	23.7	18.8	20.6	18.0	12.7	10.8	6.4	9.6	9.9	14.5	4.7	.....	2.4
{ FeSiO$_3$	10.2	8.3	9.1	7.8	11.0	9.4	7.0	10.2	8.0	9.6	8.6	12.0	5.1	.....	.....
SiO$_4$	.....	5.9	3.2	14.2	0.7	.....	.....	.....	.....	.....	.....	0.4	.....	.....	.....
iO$_4$	.....	2.8	1.3	6.6	0.4	.....	.....	.....	.....	.....	.....	0.4	.....	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.4	.....
	1.9	3.3	2.6	3.0	1.4	2.6	3.0	2.3	4.4	3.7	2.8	0.7	3.0	1.2	0.2
	2.4	2.0	1.4	1.5	1.7	2.0	1.7	2.0	4.6	2.6	2.4	2.0	2.6	0.8	1.1
	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.7	0.3	0.3	0.3
	.....	.....	.....	.....	.....	.....	.....	0.4	.....	.....	.....	.....	.....	0.9	.....
	1.0	1.0	1.1	1.0	1.7	1.4	1.0	1.2	1.1	1.4	1.1	1.2	1.6	0.8	0.5

Modes

	0.1	.....	0.1	.....	0.1	0.5	0.9	2.5	2.2	2.7	3.2	0.1	3.6	11.8	22.6
	0.1	.....	0.1	.....	0.1	0.5	0.9	1.5	2.1	2.6	3.0	0.1	2.4	.....	.....
	41.8	35.7	36.4	27.7	38.6	39.7	47.6	56.0	57.3	50.2	46.5	42.4	48.2	67.0	70.2
	50.9	52.3	50.6	46.4	56.5	56.9	48.6	37.0	35.4	38.0	41.2	49.7	32.3	17.5	4.5
	2.1	9.6	8.9	22.6	1.1	.....	.....	.....	.....	.....	.....	0.8	.....	.....	.....
	4.0	1.9	3.2	1.8	3.0	2.0	1.6	1.7	2.4	4.0	5.9	4.6	6.9	1.1	1.5
	1.0	0.5	0.7	1.5	0.6	0.4	0.4	0.4	0.6	0.8	0.2	2.3	2.5	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.7	.....	.....	4.1	.....	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.0	.....
	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	1.6	1.2

is by F. Herdsman.

Fenner (1926, p. 746-750) in his account of the Katmai Magmatic Province referred at length to the Palisade diabase. He agreed with Lewis and Bowen that the sill is an excellent example of gravity separation of crystals and pointed out that the main separation seems to have

TABLE 4.—*Analyses of Palisade Diabase from Lewis (1908a) and of rocks of similar magma type*

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	60.05	51.34	51.88	50.40	52.48	49.62	49.02	51.09	50.52	52.25	49.68
TiO_2	1.74	3.47	1.35	1.35	1.30	1.01	0.99	1.30	2.39	1.10	2.60
Al_2O_3	11.88	12.71	14.53	15.60	14.98	10.51	10.14	14.23	13.76	14.60	12.95
Fe_2O_3	3.22	2.65	1.35	3.65	1.13	0.64	1.54	2.56	3.87	0.84	3.47
FeO	10.21	14.14	9.14	6.30	9.25	12.02	10.46	7.74	8.50	9.89	10.10
MnO	0.28	0.36	0.10	0.06	0.27	0.09	0.16	0.25	0.16	0.45	0.20
MgO	0.85	3.66	7.78	6.08	7.75	15.98	17.25	7.56	5.42	6.95	5.69
CaO	4.76	7.44	9.98	10.41	10.83	7.86	8.29	10.35	9.09	9.71	10.09
Na_2O	4.04	2.43	2.06	2.57	1.87	1.40	1.59	1.92	2.42	2.21	2.27
K_2O	2.10	1.44	0.93	0.62	0.43	0.55	0.40	0.42	0.96	0.96	0.52
H_2O+	0.66	0.69	0.97	1.67	0.23	0.49	0.59	1.01	1.51	0.71	1.71
H_2O-	0.21	0.18	0.12	1.02	0.18	0.38	0.16	1.66	0.76	0.32	0.29
P_2O_5	0.52	0.20	0.14	0.16	0.13	0.16	0.11	0.16	0.26	0.22	0.33
Misc.	.....	.....	.....	.....	.....	.....	.....	.....	0.69	.....	0.04
	100.52	100.71	100.33	99.89	100.83	100.71	100.70	100.25	100.31	100.21	99.94

1. Quartz diabase, Pennsylvania railroad tunnel, Homestead about 400 feet from western portal. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
2. Quartz diabase, Jersey City, Pennsylvania railroad cut near Marion Station. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
3. Basaltic diabase, Weehawken, lower contact in Pennsylvania railroad tunnels. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
4. Basaltic diabase, New York, Susquehanna, and Western railroad tunnel, upper contact at western portal. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
5. Diabase, Weehawken, road to West Shore Ferry. Fine-grained rock in midst of olivine diabase ledge. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
6. Olivine diabase, Weehawken, road to West Shore Ferry. Typical coarse-grained olivine diabase. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
7. Olivine diabase, Englewood Cliffs. Typical coarse-grained olivine diabase. Quoted from Lewis (1908a, p. 122), R. B. Gage, analyst.
8. Watchung basalt middle "black" layer. Hartshorn's quarry. Quoted from Lewis (1908a, p. 159), R. B. Gage, analyst.
9. Average Whin Sill magma, northern England. Quoted from Holmes and Harwood (1928, p. 539).
10. Average Karroo dolerite, South Africa. Quoted from Daly and Barth (1930, p. 101).
11. Average Deccan basalt. Quoted from Holmes and Harwood (1928, p. 539).

begun and ceased "abruptly." Olivine ceased crystallizing at an early stage, and the settling of later pyroxene was impeded by the buoyant effect of feldspar and the decreasing proportions of the liquid phase. He emphasized the preponderance of quartz over olivine in the sill and considered that, while the precipitation of olivine may have increased the proportion of quartz, the greater part of the quartz would have appeared in any case. The interstitial micropegmatite of the normal diabase he regarded as probably secondary and the result of hydrothermal alteration.

Bowen (1928, p. 71-73) repeated his conviction that much of the micropegmatite formed through sinking of olivine crystals out of the upper portions of the sill; he pointed out that pyroxene had also settled, the neighborhood of the olivine diabase layer being enriched in that mineral. He considered that the olivine ceased crystallizing at an

early stage but that augite continued to form until quite late, its settling being gradually brought to an end by crowding of crystals. The complementary nature of olivine and quartz in Lewis' micrometric table was emphasized by him and also the fact that the olivine diabase showed no signs of the secondary actions advocated by Fenner. Bowen argued, in short, that crystallization differentiation alone accounted adequately for the variation in the sill.

Fenner (1929, p. 230-231, 235-236) brought forward much important evidence to show that the residual liquid of basaltic magma becomes enriched in iron and could not therefore be of granitic composition. He cited examples from Rocky Hill, which he considered to be an extension of the Palisades, and showed that the MgO/FeO ratio is higher in the pyroxene than in the corresponding rock. In the two specimens considered, that from the higher horizon is relatively impoverished in magnesia and enriched in iron. He again emphasized the importance of late hydrothermal solutions containing borates, phosphates, and halides as producers of micrographitic and myrmekitic intergrowths in diabase such as the Palisade sill.

Daly and Barth (1930, p. 109), discussing the genesis of the Karroo dolerites, drew attention to the fact that the quartz diabase of the Palisades sheet is richer in iron than the mean rock, which seemed to them to strengthen Fenner's conclusion that processes other than pure crystal fractionation had been at work. They suggested, further, that the settling of olivine was accompanied by iron oxide, the total FeO of the olivine diabase layer being greater than that of the chill phase.

Daly (1933, p. 405-406) agreed with most of Bowen's conclusions as to the differentiation of the sill but considered that the large proportion of iron in the quartz diabase still demanded explanation.

Fenner (1937, p. 159-160) recorded the presence of segregations of basic diabase pegmatite in the Palisades sill and stated that these do not have the composition to be expected in a residual liquid produced by crystal fractionation of diabase.

Butler (1937, p. 218-219) recalculated the time required for the formation of the olivine diabase layer in the Palisades sill, having observed that the average diameter of the olivine crystals is much less than the value taken by Bowen. The time of formation was found to be about seven times greater than that arrived at by Bowen. Butler also stated that differentiation was complicated by subsequent intrusions of olivine-poor diabase into the olivine-rich layer and by occasional deuteric or hydrothermal action.

Walker (1937) summarized the discrepancies recorded by previous observers and gave a preliminary résumé of the present investigation.

Bailey (1936, p. 1-6) examined the Edgewater sections of the Palisades and found that both the field and microscopic evidence supported the gravitational settling of olivine and pyroxene. He commented upon the level upper surface and irregular lower surface of the olivine diabase layer which he considered to have accumulated through the fall of a "cloud" of olivine crystals. In addition he noted the marked thinning out of the olivine diabase layer in the Edgewater road section and also suggested that there had been a rising of K_2O and SiO_2 accompanying the settling of olivine and pyroxene. He recognized that much of the pyroxene of the sill was orthorhombic.

THE CASE OF GRAVITATIONAL DIFFERENTIATION

Olivine.—It is noteworthy that all but one of the authors mentioned, though representing diverse points of view, agree that the olivine diabase layer of the Palisades was formed by the gravitational settling of early olivine. The present writer preferred to remain agnostic until an explanation had been found for certain discrepancies in the previous accounts. Detailed examination has cleared away these difficulties, and it is with relief that he now falls into line.

The discrepancies in question were as follows:

(1) The contacts of the olivine diabase layer with the normal diabase above and below were described as sometimes gradational and sometimes sharp. The first case implied a sudden cessation of the crystallization of olivine while the magma was still fluid, and the second implied a gradual halt of the settling process through increasing viscosity and crystal interference—*i.e.*, olivine continuing to separate until a later stage.

(2) If crystallization of olivine ceased at an early stage in fluid magma, it is difficult to account for the absence of packing of olivine crystals in the olivine diabase layer.

(3) The position of the olivine diabase layer is somewhat higher above the lower contact than one might expect.

(4) The normal diabase appears in places to be injected into the olivine diabase layer as hard dikelike masses.

Bailey (1936, p. 3-4) has pointed out clearly that if the olivine diabase layer did not accumulate through gravitational settling of olivine it must have been injected into almost consolidated diabase. Thus only can the mutual relationship of the two modifications and the unbroken passage of the great joint columns through the olivine diabase be explained (Pl. 1, fig. 2). It is difficult, however, to see why a relatively thin injection should persist at such a uniform distance from the lower contact throughout its 20-mile outcrop, even following it where it transgresses. Bailey has stated as evidence against the injection theory that the lower boundary of the olivine diabase layer is highly irregular, while the upper boundary runs as an almost straight line, and that in the Edgewater road cut the

irregularity of the lower boundary is so marked that at one point the upper diabase rests directly on the lower, to the exclusion of the olivine diabase layer. The author agrees that the upper boundary is more regular than the lower, though by no means a straight line, and records an apparent local thinning of the olivine diabase layer to 3 feet (Pl. 1, fig. 1) but he has not observed its total disappearance in the Edgewater road cut or in any other section.

Close examination in the field shows that the boundaries of the olivine diabase layer are generally much less sharply defined than they appear to be when viewed from a distance, and microscopic examination of thin slices provides the explanation. It is then found that specimens, which appear in the field to be outside the boundaries of the olivine diabase layer, are identical with those from within. This seems to be due to differential weathering of the olivine diabase layer. As a rule it weathers (through expansion of partly serpentinized olivine crystals) to a crumbly surface which, unlike the smooth joint surfaces of the normal diabase, reflects little light. Thus it shows as a dark band along cliff sections (Pl. 1, fig. 2). If, however, portions of it remain unaltered, as frequently happens, it is exceedingly difficult to distinguish the resulting smooth surfaces of olivine-rich diabase from the adjacent modifications. Thus the apparent thinning of the olivine diabase layer at the Edgewater road cut may not be an actual one.

Examination of the micrometric tables shows a somewhat gradual upward increase of the olivine percentage from the lower contact. This increase becomes much more rapid in the immediate vicinity of the olivine diabase layer where a maximum is reached. Above this basic layer there is an equally rapid decrease, and olivine is not found more than 20 feet above it in any section. Though these increases and decreases of the olivine percentage are rapid, it cannot be emphasized too strongly that they are gradational and that no discontinuity was found in any section. Sections 4, 5 (Fig. 5), and 6 bring out this point clearly. It will be remembered that the olivine diabase layer contains very numerous small crystals of olivine which are not found elsewhere and that these are distinctly richer in fayalite than the larger crystals. The above facts do not accord with the theory of subsequent injection but may be explained by a hypothesis of gravitational settling as follows.

Crystallization of the magma commenced with the separation of olivine,[1] probably in the metastable region, resulting in the fall of relatively few large crystals into the lower chilled phase which was growing upward. They became incorporated among the pyroxene and plagioclase crystals

[1] The basaltic contacts contain 3 to 5 per cent of phenocrysts (plagioclase, augite, and olivine) showing that crystallization had commenced before emplacement of the sill.

of this phase above levels varying from 10 to 30 feet from the lower contact but were insufficient in numbers to show packing and therefore have a somewhat scattered distribution. Separation of olivine probably ceased in the metastable region in the upper parts of the sill, which had become relatively siliceous by the abstraction of the orthosilicate. It is thought that in the lower portion of the sill, which now contained a concentration of orthosilicate, separation of olivine continued into the labile region, when a shower of small crystals (richer in fayalite) was precipitated. These crystals fell in sufficient numbers to show signs of packing between the larger constituents (Pl. 2, fig. 6) and accumulated to form the olivine diabase layer. Separation of olivine then ceased, but the fall of small crystals was accompanied by larger crystals of olivine, which had made their way down from high levels in the sill and continued to arrive for a short time after the formation of the olivine diabase layer, to become incorporated in the diabase immediately above it.

The explanation differs in some respects from the one advanced by Bailey. No evidence was found for the enclosure of olivine in pyroxene alone below the olivine diabase layer, and preferentially in plagioclase above it, with no preference in the layer itself, as recorded by him (Bailey, 1936, p. 5-6, Fig. 1).

The following figures were obtained for slices from the Edgewater road cut and do not include a great number of olivine crystals which were approximately half in plagioclase and half in pyroxene:

	No. of olivines in pyroxene	No. of olivines in plagioclase
18 inches above the olivine diabase layer	54	36
2 feet below the top of the olivine diabase layer	191	31
10 feet below the olivine diabase layer	60	43

Since plagioclase and pyroxene occur in roughly equal proportions, it may be concluded from these figures that the former was in general the first to crystallize.

A point which is definitely in favor of gravitational settling of olivine and against subsequent injection of the olivine diabase layer is the fact that the pyroxene-plagioclase ratio in the layer itself, and in the modification above and below, remains fairly constant. Such variations as do occur show no regular tendency. This is seen in Table 5 which gives figures for sections (4) Edgewater Cliff and (6) Hoboken.

A feature common to all the sections except (1) is the marked increase in size of the plagioclase crystals in the olivine diabase layer over those in its vicinity (Fig. 3). It is somewhat difficult to explain, but the author has noticed a similar phenomenon in two Scottish sills (Walker, 1930b, p.

365; 1932, p. 246) in the portions showing gravitational concentration of olivine. In these cases, however, the original magma had the composition of olivine basalt, and the basic phase consisted of picrite. It would seem that the large size of the plagioclase crystals is in some way connected

TABLE 5.—*Ratios of pyroxene to plagioclase*

Edgewater Cliff (4)			Hoboken (6)		
Height above contact (Feet)	Pyroxene-plagioclase ratio		Height above lower contact (Feet)	Pyroxene-plagioclase ratio	
80	0.92		80	0.80	
77	0.79		79	0.91	
74	0.88		78	0.80	olivine
72	0.88		70	0.81	diabase
70	0.81	olivine	60	0.81	layer
68	0.74	diabase	59	0.94	
63	0.87	layer	52	0.83	
61	0.69		48	0.78	
60	0.96				
57	0.86				
35	0.84				

with concentration of orthosilicate. In the Alpine section where the olivine diabase layer is much less marked the feldspar is of normal size.

The unexpectedly high position of the olivine diabase layer in the sill is to be explained by the fact that this layer represents the last crystallization of olivine at a time when the marginal solid phase had been able to grow upward for some distance.

Most of the hard dikelike bands in the olivine diabase layer are coincident with joint cracks which acted as conduits for hydrothermal and metasomatic solutions. It is thought that the hard bands arose through the toughening and cementing action of these solutions, the former of which left traces of their passage by precipitation of biotite. (An alternative explanation of similar phenomena in the basalts of Oregon is given by Fuller (1938).) Other hard bands in the basic layer demand a different explanation. Lewis (1908a, p. 126; Pl. 20, fig. 2) has described and figured dikelike apophyses of fine-grained diabase in the olivine diabase layer containing only a small proportion of olivine and gives an analysis of one from the West Shore Ferry road at Weehawken.

The author has not seen this example, but at the neighboring locality of King's Point he encountered one of the very rare discontinuities of crystallization in the lower part of the sill. About 4 feet below the olivine diabase layer there is a stratum of almost basaltic diabase which has much the same mineralogical composition as the adjacent rock but is considerably finer in grain. The contact between the two facies is sharp and unchilled. A similar discontinuity in exactly the same position was found

in the core of a bore sunk near by in connection with the Midtown Tunnel and made available to the author through the kindness of the engineering authorities. It is suggested that these fine-grained rocks represent subsequent intrusions of virtually undifferentiated magma into the solid rock of the sill while it was still hot. The phenomenon is by no means an

TABLE 6.—*Chemical analyses*

		SiO_2	Al_2O_3	Fe_2O_3 +FeO	MgO	CaO	Na_2O	K_2O
(A)	Basalt intrusion	52.5	15.0	10.4	7.8	10.8	1.9	0.4
	Contact basalt	51.9	14.5	10.5	7.8	10.0	2.1	0.9
(B)	Basalt	50.1	14.9	12.8	5.4	8.8	2.1	1.3
	Near-by dolerite	49.9	14.6	12.9	5.9	8.7	2.8	0.7
(C)	Basalt vein	48.1	15.2	13.5	5.6	9.9	2.3	0.6
	Adjacent dolerite	50.0	15.0	12.9	4.2	9.2	2.4	1.2

(A) Weehawken (Lewis, 1908a, p. 121, analyses 4 and 5).
(B) Binny Craig (Lunn, 1928, p. 75, analyses 1 and 2).
(C) White Force (Smythe, 1930, p. 103, analyses 4 and 5).

isolated one and demands further investigation. It has been recorded in Permo-Carboniferous quartz diabases of tholeiite magma type in the south of Scotland (Lunn, 1928, p. 74-79) and the north of England (Smythe, 1930, p. 102-107), the mutual relations of the two facies being essentially the same. It is also noteworthy that in the three cases where comparisons may best be made the chemical relationship of the two facies is similar. This is brought out clearly in Table 6. In the Palisade group (A) the composition of the early undifferentiated magma is represented by the lower basaltic contact for this alone has escaped differentiation. In the Scottish example (B) no analysis of the "dolerite" which is undifferentiated, as is that in (C), is available so that of the nearest undifferentiated quartz dolerite sill is substituted.

Pyroxene.—The occurrence of pyroxene phenocrysts in the basaltic contact modifications shows that the mineral had begun to crystallize before the magma was emplaced. Its solidification *in situ*, however, was probably not rapid during the precipitation of olivine, for olivine seems to have sunk unimpeded and exhibits idiomorphic boundaries toward all varieties of pyroxene. On the other hand, it seems likely that the crystallization range of the two minerals overlapped to some extent, for there is a concentration of pyroxene in the lower part of the sill commencing considerably below the olivine diabase layer. Such a concentration could be caused only by the gravitational settling of pyroxene crystals to a growing lower chilled phase aided to a certain extent, perhaps, by reaction of liquid magma with sunken olivine crystals.

The proportion of pyroxene (except in section 6) shows a distinct upward increase below the olivine diabase layer but falls off in and adjacent to it where the percentage of olivine is high, the pyroxene-plagioclase ratio remaining approximately constant. Above the olivine diabase layer the increase in the proportion of pyroxene continues until a maximum is reached 70 to 100 feet above the base of the sill. Above this level there is a gradual falling off in the percentage up to near the upper chilled phase. It is clear that the concentration of pyroxene has been caused by gravitational settling which gradually ceased because of crystal interference and increasing viscosity of the magma. Crystallization of pyroxene was accompanied by that of plagioclase, the metasilicate often enclosing laths of plagioclase which must have been carried down with it.

Assuming an intrusion temperature of 1100° C. (Sosman and Merwin, 1913, p. 394) for the magma, and using the thermal expansion data of Day, Sosman and Hostetter (1914) and of Kozu and Ueda (Kozu, Ueda, and Tsurumi, 1934, p. 83-86; Kozu and Ueda, 1934, p. 87-90), the specific gravity of the magma at the time of emplacement was calculated to be 2.64, while the values for the early olivine and pyroxene at the same temperature were found to be 3.27 and 3.23 respectively. The density differences between minerals and magma (0.73 and 0.59) seem quite sufficient to cause crystal settling, and, while the difference between pyroxene and magma is the smaller, the larger size of the pyroxene crystals would tend to compensate for this.

The suggestion of Daly and Barth (1930, p. 109) that the sinking of olivine was accompanied by iron ore cannot be entertained. The percentage of iron ore in the normal rock of the sill is low, particularly in the olivine diabase, and tends if anything to show a decrease downward.

It is also very unlikely that plagioclase crystals rose or fell in the magma to any appreciable extent. The specific gravity of the early plagioclase at 1100° C., extrapolating from the data of Kozu and Ueda (1933, p. 262-264), would be 2.65 which is actually very slightly higher than the value obtained for the magma at that temperature.

Lastly there is always the possibility that the two-phase convection process of Grout (1918, p. 491-499) operated to some degree, though no evidence can be adduced to support this. The recent work of Nichols (1939, p. 293) on the viscosity of basalt lavas would seem to indicate that the viscosity values taken by Bowen and Butler in their calculations of the time of formation of the olivine diabase layer were probably much too low. It follows then that not only are the times given by them too short but also that the likelihood of convection currents operating to any great extent is correspondingly reduced.

When the specific gravity of the diabase at various levels in sections 3 and 4 is plotted against the height above the base of the sill (Fig. 7) it becomes obvious that gravity has played a great part in the differentiation. The marked rise above the average value (as shown by the contacts) in the lower part of the sill, the complementary fall below it in the

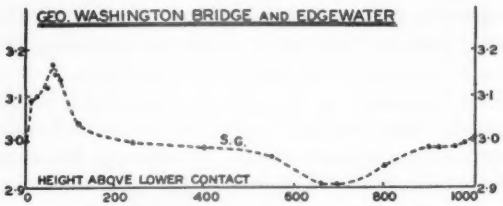


FIGURE 7.—*Specific gravities of sections*
At George Washington Bridge and Edgewater.

upper part, and the absence of any discontinuity can lead only to this conclusion.

CRYSTALLIZATION OF THE PYROXENES

In recent years our knowledge of the pyroxenes has been greatly increased by the fine experimental work of Bowen and Schairer on the relevant silicate systems, and by a number of valuable papers by American, British, and Japanese petrologists, correlating optical properties with chemical composition in naturally occurring examples. Some agreement has also been reached as to the part played by the group on the crystallization of basaltic magma, though there still remains considerable divergence of opinion on this. In this connection the pyroxenes of the Palisade sill are interesting.

It seems to be generally agreed that the groundmass pyroxenes of effusive rocks of tholeiite magma type are monoclinic, characterized by low axial angles, and are zonal, the marginal portions being lower in axial angle and richer in iron than the centers. Phenocrysts, when present, tend to be diopsidic with larger axial angles, but published accounts indicate that basalts of this magma type are typically nonporphyritic. In some basalts there seems to be a continuous gradation of 2V between 50° and 0°. Many dolerites of equivalent chemical composition show the same features, notably those of the Karroo (Daly and Barth, 1930, p. 97, 110).

Under deep-seated conditions, however, it appears that pigeonitic pyroxene is unstable, its place being taken by orthopyroxene and a diopsidic clinopyroxene. There is, in short, evidence of limited miscibility of the diopside-hedenbergite with the clinoenstatite-ferrosilite series. The boun-

dary of instability occurs within the Palisade sill, as is evident from the descriptive mineralogy of its pyroxenes. In the upper two-thirds of the sill all the pyroxene is monoclinic and mainly of low axial angle; in the lower third the clinopyroxene is practically all of relatively high axial angle (mainly 45° to 50°) and is accompanied by abundant hypersthene. Much of the hypersthene shows the effects of unmixing and contains graphic intergrowths of lime-rich clinopyroxene. The marginal portions of the crystals, however, lack these inclusions, indicating that the final crystallization of pyroxene was almost free from lime.

The instability boundary of pigeonite is thus located between 300 and 400 feet above the lower contact, but the reason for its location is doubtful. It cannot be a case of temperature, for, although Bowen and Schairer (1935, p. 153-217) have shown that hypersthene is the low-temperature form of pigeonite, there is no reason to believe that the crystallization of the Palisade pyroxenes took place at lower temperatures beneath the instability boundary than above it. The stability of the pigeonitic pyroxenes above this boundary may be due to a relative poverty of magnesia or to a relative abundance of volatiles. The author is inclined to favor the latter for in samples collected from coarse schlieren of the English Whin Sill—a typical quartz diabase—he has, like Holmes (1928, p. 504) and Tomkeieff (1929, p. 106-108), encountered only clinopyroxene. Much of this pyroxene is pigeonitic, and, as Tomkeieff (1928, p. 117-118) has pointed out, these coarse schlieren represent "wet" fractions of the magma. The "dry" fractions, which are of finer texture but similar chemical composition, frequently contain orthopyroxene as well as hypersthene-augite. It is, at any rate, clear that the Palisade pigeonite is not a rapidly cooled metastable phenomenon, for the coarseness of crystallization completely precludes this explanation though it favors a concentration of volatile constituents.

The order of crystallization of the pyroxenes presents some interesting features. The Palisade diabase provides one of the rare cases in which the orthopyroxene is of late formation. The normative pyroxenes (reduced to 100) of the hypersthene-bearing types of Table 3 should therefore fall above Tsuboi's (1932, p. 75) provisional two-pyroxene boundary (CD on Figure 8) and this they do without exception. Above the instability boundary of pigeonite the sharply defined uniaxial cores of some crystals demand explanation. They possibly represent completed reaction between magma and early formed olivine crystals, as Barth (1936, p. 330) has suggested in the case of similar effusive types. Such cores are also found, though very rarely, in the clinopyroxene of the lower part of the sill. Elsewhere the crystallization of the pyroxene has followed normal lines from

hypersthene-augite toward iron-rich pigeonite, though zoning is not very conspicuous.

It will be observed in Figure 8 that all the Palisade rocks carrying both ortho- and clinopyroxene lie within Asklund's (1925, Figs. 20 and 21)

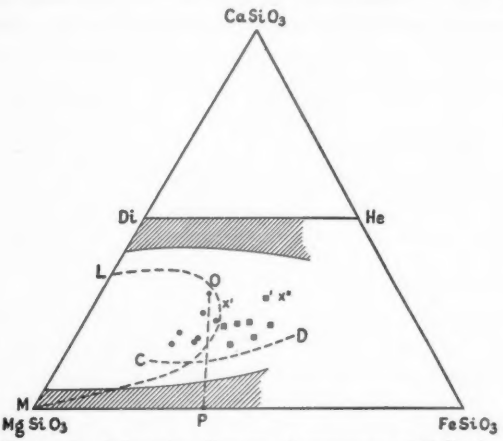


FIGURE 8.—*Triangular diagram of pyroxenes of Palisade sill*

⊙ = normative pyroxene (Table 3) in rocks with two pyroxenes.
⊡ = normative pyroxene (Table 3) in rocks with clinopyroxene only.
⊡′ = normative pyroxene (Table 3) in diabase pegmatite.
×′ = pyroxene from Rocky Hill, middle of sheet.
×″ = pyroxene from Rocky Hill, upper part of sheet (diabase pegmatite). ×′ and ×″ derived from analyses No. 2 and 3 (Phillips, 1899, p. 281).
Upper and lower hatched areas represent stability fields for clino- and orthopyroxene respectively in rocks with two pyroxenes (Tsuboi, 1932).

boundary for rocks with two pyroxenes (L-M) and all of those with only clinopyroxene lie outside it. The normative pyroxene of the pegmatite differentiate also closely approximates in composition the analyzed pyroxene from the upper part of the Rocky Hill sill (Phillips, 1899).

It is known that the final product of pyroxene crystallization in the lowest third of the sill is hypersthene with about 40 per cent ferrosilite (P. on Figure 8). The composition of the early clinopyroxene O has also been deduced, and the course of crystallization is from O and P toward $FeSiO_3$. It is satisfactory to note that the normative pyroxenes of all the rocks in this part of the sill lie, as might be expected, roughly between O and P. The position of O is not fixed with any great confidence in its accuracy for, as Deer and Wager (1938, p. 21) have pointed out, small amounts of

minor constituents such as TiO_2 may have profound effects on the optical properties of clinopyroxene, and the composition of O may well be nearer the wollastonite-clinoenstatite side of the triangle. It seems clear in any case that the upper hatched area which, according to Tsuboi, represents the stability field for clinopyroxene in two pyroxene rocks should be extended downward, for the Palisade clinopyroxene lies far outside it.

LATER STAGES OF DIFFERENTIATION

Crystal settling in the Palisade sill led to an accumulation of the solid phase in the lower parts while the upper parts remained fluid. Since all the volatile constituents remained in the liquid phase, as crystallization progressed they became concentrated increasingly in the higher levels below the upper chilled phase. Such a process would be brought about by normal crystal fractionation alone, as emphasized by Koomans and Kuenen (1938, p. 156-157) in the case of the Glen More ring-dike of Mull and need not involve any other upward migration of volatiles. Richness in volatiles tends to promote coarse crystallization, and the variation in grain size from floor to roof of the sill is quite in keeping with the process outlined, if the pegmatite schlieren are excepted. These schlieren afford evidence of a mobile residue which may have migrated independent of the normal fractionation process. They have an irregular banded or patchy distribution and variable composition which seem incompatible with the regular and progressive enrichment in silica of the residual fractions which should be produced by crystal settling and may be observed in the remainder of the sill.

Such schlieren are exceedingly common in the upper parts of quartz diabase sills from all over the world and are recorded in examples too thin to show gravitational differentiation. The illuminating study of the Whin Sill by Tomkeieff (1929) proves that they are "wet" fractions quite close in composition to the normal rock, though usually slightly richer in iron and alkalies and poorer in magnesia. The presence of volatiles has, however, entirely changed the mineral composition by preventing iron oxide from going into chemical combination with silica as pyroxene. The pegmatitic rock is thus rich in quartz and late iron ore. This tendency for the volatiles in diabases to hold back iron oxide and expel it as iron ore or chlorophaeite, just before the last stages of crystallization, has been observed by Campbell and Lunn (1928, p. 500) and by the author (Walker, 1930a, p. 374-376) in Scottish rocks.

The pegmatite schlieren of the Palisade sill show a similar enrichment in quartz and iron ore (Table 2), and the figures quoted by Lewis for coarse-grained quartz diabase from Homestead and Marion (Table 4, 1 and 2) are extreme cases. The second rock, with 16.79 per cent total iron

oxide, evidently represents a residual liquid segregated just before the iron ore was precipitated from solution and contains 17 per cent of ores associated with only 7 per cent of quartz. The importance of this process in determining the mineralogy of tholeiitic rock types has not been fully realized. In the case of the first rock we have probably the segregation of a residue from which a good deal of iron ore had already crystallized, for 19 per cent of quartz is associated with only 6 of ore. It is plain that the residual liquid which gave rise to the pegmatite schlieren was mobile, rich in iron, and of very variable composition.

The formation of the pegmatite schlieren probably ushered in the deuteric stage of crystallization. The plagioclase which they contained was fairly calcic, usually An_{45-50}, and this began to suffer albitization by the volatile-rich residual magma. To this stage, too, belongs the conversion of much of the pyroxene to green or brown amphibole, the formation of biotite and sodic amphibole by reaction with iron ore and pyroxene, and also the formation of a good deal of quartz and micropegmatite.

The deuteric stage led gradually into the hydrothermal when the calcic cores of some of the plagioclase crystals were replaced by micropegmatite and pectolite was formed. The hot solutions, rich in water, silica, and alkalies, percolated through the upper chilled phase depositing some poikilitic biotite, probably by reaction with titanomagnetite, and some quartz and alkalic feldspar. Similar solutions seem to have been present in the lower part of the sill, but they were doubtless far less abundant than in the upper portion; biotite formed here by reaction with the mafic constituents of the diabase, mainly near joint fissures which acted as conduits. Even in unaltered diabase below the olivine diabase layer, and containing as much as 6 per cent of olivine, there is a small proportion of interstitial micropegmatite showing that there was a siliceous residue.

In both cases the residual liquids seem to have freed themselves from potash almost entirely, with the result that they became highly sodic and gave rise to the quartz-albite-pyroxene veins, which are found in both the upper and lower chilled phases and represent the expulsion of the last magmatic residues. These sodic veins were undoubtedly injected into solid diabase and have sometimes caused a certain amount of chloritization and serpentinization in their immediate vicinity. They find counterparts in diabases from Scotland, Sweden, and Virginia as is seen by the analyses in Table 7.

The first five analyses in Table 7 are highly sodic and very low in potash. Although they are similar in most respects the proportion of lime is very variable. The last two analyses are potassic but otherwise resemble the first five (No. 6 contains 8.75 per cent H_2O). No. 4 is characterized by prominent graphic intergrowth between quartz and albite, but the tex-

ture of the others is felsitic. Where the ferromagnesian mineral is fresh it proves to be augite—a pale-green soda-bearing variety in the case of No. 5. It is thought that the first five rocks represent the final magmatic residuum and verge on the deuteric or even the hydrothermal stage of crystallization.

TABLE 7.—*Chemical analyses of pegmatite schlieren and residual injections*

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
(1)	71.60	13.16	1.28	0.38	2.12	3.76	5.92	0.70
(2)	68.74	13.24	1.22	1.38	2.02	5.90	5.76	0.36
(3)	71.26	11.87	0.10	2.12	1.08	2.88	6.73	0.05
(4)	70.75	12.01	0.62	0.94	0.56	6.13	7.20	0.79
(5)	72.52	11.84	1.23	0.87	1.48	2.54	7.12	0.36
(6)	66.80	12.10	0.97	1.50	0.50	2.62	2.40	4.20
(7)	72.70	13.20	0.37	2.29	0.55	1.40	2.19	5.15

(1) Diabase aplite, Goose Creek, Virginia (Shannon, 1924, p. 28).
(2) Segregation vein in diabase, Goose Creek, Virginia (Shannon, 1924, p. 25).
(3) Segregation vein in quartz dolerite, W. Lothian, Scotland (Falconer, 1906, p. 147).
(4) Augite-albitophyre in diabase, North Ulfö, Sweden (Von Eckermann, 1938, p. 420).
(5) Diabase aplite, George Washington Bridge, Palisades (From Table 3).
(6) Residual glass from tholeiite, Kinkell, Scotland (Walker, 1935, p. 150).
(7) Felsitic rock in Whin Sill, Dunstanburgh, England (Smythe, 1930, p. 103).

Analyses 6 and 7 show considerable resemblance but belong to the previous stage—before the removal of the potash as alkalic feldspar and biotite—in which no hydrothermal alteration can have taken place. If the interstitial material from the normal diabase of the Palisades were isolated it would probably be found to have much the same composition as Nos. 6 and 7.

The author believes that the elimination of iron oxide and then potash from the magma leaves a highly sodic residuum, which is to be found veining many diabases of tholeiitic magma type. He is at present attempting to collect further evidence in support of this hypothesis and considers further that albite is the only stable feldspar under very late magmatic or hydrothermal conditions; any lime present at that stage tends to form diopsidic pyroxene or pectolite rather than soda-lime feldspar. This is borne out by the white veins of the Palisades, which consist of a mosaic of quartz, albite, colorless to pale-green augite, and also by similar veins described by Von Eckermann (1938, p. 420) from Swedish diabases and by Shannon (1924) from Goose Creek, Virginia.

The analyses of the pegmatite schlieren and residual injections (Table 7) show great variety of chemical composition. The former differs only slightly from the composition of the normal diabase at the same horizon and from these stages may be traced to the extreme sodic differentiate. This variation is connected, in the author's opinion, with the nature of the zoning in the chief constituents of the magma—*viz.*, pyroxene and plagio-

clase. In each case the zoning from the center outward is progressive but comparatively slow until a narrow peripheral area is encountered in which the change of composition becomes much more rapid. Thus, after the abstraction of accumulation by gravity of olivine and pyroxene, the composition of the solid and liquid phases in the normal diabase showed little

TABLE 8.—*Comparison of chemical analyses of diabase and associated diabase pegmatite*

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
(1)	51.46	16.24	3.04	9.58	3.47	8.38	3.01	0.87
(2)	52.50	18.28	1.97	7.61	2.58	9.34	3.78	1.03
(3)	50.51	14.61	3.50	8.85	5.32	8.76	2.55	1.19
(4)	49.70	15.64	2.42	11.35	3.06	8.03	2.90	1.22
(5)	51.56	13.81	0.96	11.32	7.40	10.08	2.08	0.96
(6)	52.94	14.80	0.16	12.00	5.42	8.32	1.98	1.50

(1) Diabase, Palisades.
(2) Diabase pegmatite, Palisades.
(3) Diabase (average), Whin Sill (Tomkeieff, 1929, p. 106).
(4) Diabase pegmatite, Whin Sill (Tomkeieff, 1929, p. 106).
(5) Diabase, Goose Creek, Virginia (Shannon, 1924, p. 13).
(6) Diabase pegmatite, Goose Creek, Virginia (Shannon, 1924, p. 19).

variation until the liquid phase represented perhaps 10 per cent of the whole. Residual pegmatite liquid isolated at any stage until then would differ only slightly in composition from the normal rock. During the crystallization of the last 10 per cent or so of magma, however, there were very marked changes in the composition of both solid and liquid; the pyroxene became enriched in iron, and the plagioclase in alkalies. Consequently the liquids isolated during these last stages (which are characterized by the separation of much iron ore) might show a very considerable range of composition, which is reflected in the varied mineralogy of the schlieren and residual injections which are their representatives. In the quartz diabase sills of England (Whin Sill) and of Goose Creek, Virginia, the pegmatite schlieren have a very similar chemical relationship to the normal diabase; this is brought out clearly by the analyses in Table 8.

In each case the diabase pegmatite, though similar in composition to the normal rock, is slightly more salic and is characterized by higher alkalies and alumina and distinctly lower magnesia. It represents the segregation of liquid at a relatively early stage. Analyses 1, 2, 5, and 7 in Table 7 from the same sills give the composition of the late extreme differentiates.

In the Palisades sill, and indeed in nearly all differentiated quartz diabase sills,[2] the pegmatite schlieren and residual injections bulk most prominently in the upper parts. They are by no means absent, however, from

[2] The Goose Creek diabase (Shannon, 1924, p. 2-3) is apparently an exception, but the field relations are not well exposed. The author has visited this locality and is also familiar with the coarse modifications of the Whin Sill.

the lower parts, though the diabase pegmatite is confined to thin veins, which have mineral composition identical to the most basic of the schlieren near the top.

It is clear that the white aplitic veins were injected into solid diabase, but the mode of emplacement of the pegmatite schlieren remains in doubt. They are obviously connected with the migration and local segregation of volatiles, and some of them, at least, seem to be the result of auto-intrusion either into pasty magma or into rifts in the crystal mesh.

THE CASE AGAINST SYNTEXIS

Examination of Lewis' (1908a, p. 123) micrometric data seems to indicate that in the sill, as a whole, quartz is far more abundant than olivine and that the percentage, if averaged over the whole intrusion, would be about 6. Since the norm of an analysis of the quenched lower contact contains only 2 per cent of the mineral, there would appear to be a good case for syntexis of diabase with arkose, particularly as considerable slabs of arkose have been demonstrably floated up from the floor. Lewis, however, provides convincing field evidence against syntexis, and the additional chemical and micrometric data now brought forward show that syntexis was not important.

Of the numerous upstoped slabs of arkose recorded by Lewis, one at Weehawken demands special attention. It has been floated into a vertical position and now extends from the top to the bottom of a cliff face. The lowest visible point is probably about 100 feet above the base of the sill, which is concealed by talus. The slab has thus been carried up into the slowly cooled central portion of the intrusion, where there was abundant opportunity for reaction with the magma. It is found, however, that the arkose inclusion has sharply defined boundaries and actually retains, as Lewis (1908a, p. 135) has recorded, a diagonal lamination representing the cross-bedding, which is a characteristic feature of the Newark formation. Slices were made of specimens from the center and the margin of the slab, and of the diabase at its contact with the arkose and 1 foot away from the contact. There was no visible silication of the diabase in the hand specimen, and the two slices of diabase proved to be identical, micrometric measurements giving the following results:

	Pyroxene	Plagioclase	Iron Ore	Biotite	Quartz and micropegmatite	Average length of plagioclase
Diabase at arkose contact.....	45.9	49.8	1.5	0.6	2.2	0.8 mm.
Diabase 12 inches from contact...	45.4	49.3	1.9	1.0	2.4	0.8 mm.

The arkose shows a slight development of micropegmatite at the contact but is otherwise unchanged.

This is believed to be very strong evidence against syntexis, and the chemical and micrometric data seem to complete the case. If the normative quartz of the normal diabase (Table 3) is averaged over the sill, and the extra quartz of the pegmatite schlieren and residual injections considered to be balanced by the olivine in the lower parts of the sill, the average normative quartz content is found to be just lower than that of the lower quenched contact (2.15 against 2.5).[3] These figures agree almost exactly with the average modal quartz content of the very complete Englewood and George Washington Bridge sections (2.0 and 1.7).[4] There is accordingly no evidence for syntexis or introduction of silica into the sill.

Although the author's slices of included arkose show few signs of recrystallization, Lewis has described examples which he considers to have recrystallized with addition of magmatic iron oxide to form a "recomposed augite-granite." It is evident, however, from the accompanying analysis, that the magmatic addition was slight.

Inclusions of shale have also been stoped up by the magma (*e. g.*, below the George Washington Bridge) and show much more complete reconstitution with development of considerable alkalic feldspar. They are not found above the lower chilled phase for they are probably denser than the arkose and would not float. The calculated densities of diabase and arkose at 1100°C. are 2.64 and 2.50, respectively.

Some of the arkose immediately below the diabase at Coytesville has apparently been affected by sodic solutions issuing from the igneous rock for it contains granules of green soda pyroxene.

The white veins which penetrate the chilled phase of the sill have a microgranitic texture quite similar to that of the baked arkose, and the author therefore considered the possibility of their having a rheomorphic origin—*i. e.*, that they represent arkose mobilized by the igneous rock and then injected into it. Detailed examination showed, however, that feldspar of the veins is albite without exception, while a good deal of potash feldspar is found in the arkose whether included in the diabase or adjacent to it. There are, in addition, slight textural differences which are brought out (Pl. 2, figs. 1 and 2). The differences of chemical composition between the highly sodic veins and the arkose are seen clearly in Table 9.

The behavior of arkose slabs included in the diabase and the observations of Sosman and Merwin in this connection are completely opposed to a rheomorphic origin for the white veins. Such veins appear to be a normal sodic differentiate of diabase magma, whatever the nature of the

[3] Actually it would be more than balanced by the olivine.

[4] This was determined by graphical and planimetric methods from Table 1. If the constituents of the analyses are similarly treated a good match is obtained between the average composition of the sill and the composition of the lower quenched contact.

country rock. The only puzzling feature is the occurrence in one slice of a white vein from the Englewood Cliff diabase of two large corroded crystals of quartz. These, however, are no larger than some of the quartz crystals in the pegmatite schlieren.

The author wishes to emphasize that the case made out by him against syntexis applies to the Palisade diabase alone. Mountain (1936, p. 93-112) has described one of the Karroo dolerites which has apparently been acidified by arkose inclusions and injected by rheomorphic veins of arkose, but the subject clearly demands further investigation.

TABLE 9.—*Chemical analyses of arkose and white veins in diabase*

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
(1)	72.62	13.86	0.31	0.44	1.24	1.01	5.57	3.58
(2)	74.99	10.96	0.36	2.70	1.37	1.80	4.37	2.21
(3)	72.52	11.84	1.23	0.87	1.48	2.54	7.12	0.36

(1) Arkose 3 feet below lower contact, Coytesville (Lewis, 1908a, p. 137).
(2) Arkose inclusion in diabase, Marion, Jersey City (Lewis, 1908a, p. 137).
(3) White vein in diabase 20 feet above contact, Coytesville (Table 3).

CHEMICAL ANALYSES

Reference has already been made to the chemistry of the Palisade rock types, and the generally satisfactory tally of norm and mode in Table 3 has been remarked upon.

It may be observed, however, that modal pyroxene is consistently greater than normative. This is probably due to the occurrence in the modal pyroxene of constituents such as alumina and titania which are not included in the normative metasilicates. In the case of the white vein the normative quartz greatly exceeds the modal, and no adequate explanation can be offered, though there is certainly some variation in the mineral constitution of these injections. It is perhaps a little surprising to find acmite among the normative minerals, but the pyroxene of the rock has a greenish tinge suggestive of the presence of soda.

Fenner (1926, p. 713) has shown that Gage's Palisade analyses when plotted in a normal variation diagram with silica as the abscissa and the other oxides as the ordinate produce a greatly scattered array of points through which no curves can be drawn. The new analyses of Table 8 accentuate this scattering for which the following explanation accounts. The main differentiation of the sill has been brought about by the sinking of a constituent (pyroxene) whose silica percentage is practically the same as that of the magma. Hence the great majority of the rocks in the diagram, although widely different chemically, are crowded into the narrow area between silica percentages of 48 and 52, leading to a quite unintelli-

gible distribution of points. Experiments have been made with other types of diagrams including those devised by Larsen (1938), but in no case with satisfactory results. The Palisades analyses seem to yield quite as much information from inspection of the tabulated figures as from any

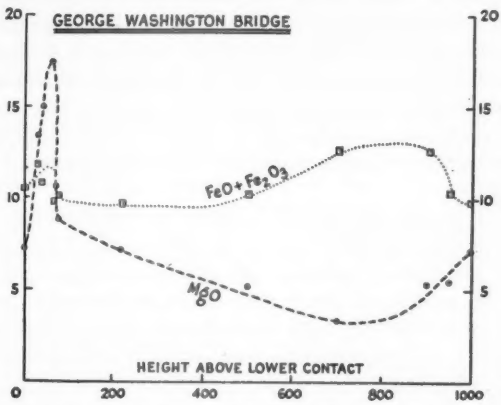


FIGURE 9.—*MgO and FeO + Fe_2O_3 plotted against height above lower contact of Palisade sill*

graphical presentation thereof. If, however, the total iron and magnesia of the new analysis of diabase are plotted against the height above the lower contact (Fig. 9), it will be seen clearly how magnesia-rich minerals with a moderate percentage of iron oxide have accumulated above the lower chilled phase, leading to an impoverishment in these constituents in the central portions of the sill. The high percentage of iron oxide in the upper part is noteworthy and doubtless indicates an upward migration of that constituent in solution.

DIFFERENTIATION OF SIMILAR INTRUSIONS

Although differentiation in some degree is a common feature of large intrusions of tholeiitic magma type, few are really comparable with the Palisades sill. A common feature is the segregation of a coarse, "wet," granopyhric facies below the upper chilled phase of sills, but the accumulation of a lower basic zone, by gravitational settling of mafic constituents, is much rarer. Of the relatively small number of intrusions of this type which do show unmistakable gravitational differentiation only very few have been described in detail with adequate optical, micrometric, and chemical data. One of the most interesting of these which have been described in detail is the Insizwa-Ingeli sheet of East Griqualand and Pondo-

land described by Du Toit (1910, p. 114-118) and later, with a wealth of optical and micrometric data, by Scholtz (1936, p. 82-210). This enormous intrusion, which represents typical Karroo Dolerite magma, similar to that of the Palisades, is 3000 feet thick and is markedly transgressive, forming a series of concavo-convex lenses. It exhibits very thorough gravitational differentiation, the roof zone consisting of granite or quartz monzonite while the basal zone is picrite. There is a tendency for the basal zone to thicken in the troughs and for the roof zone to thicken in the ridges. It is difficult to see how this thickening arose unless there is a definite upward migration of acid material, apart from crystal fractionation, and this is an argument for the same process playing a part in the Palisades differentiation.

The Glen More ring-dike of Mull (Bailey and others, 1924, p. 320-330) is fully described in the Mull Memoir. It has recently been reinvestigated by Koomans and Kuenen (1938, p. 145-160), who stress gravitational differentiation by settling of augite, iron ore, and plagioclase, without upward migration of acid material as suggested in the Memoir. In this example neither roof nor floor is seen, but the differentiation is very thorough, varying from granophyre to gabbro rich in iron ore and pyroxene. Little, if any, olivine crystallized, and the initial magma had somewhat higher alkalies and lower lime and magnesia than that of the Palisades.

In both the above cases differentiation has been much more thorough than in the Palisades. The sinking in quantity of such constituents as olivine and iron ore has both lowered the silica percentage near the bottom and raised it near the top so that the percentage range of this constituent is much greater than in the Palisades where only a small proportion of olivine sank.

A more analogous case is that of a 556-foot tholeiitic lava flow from Cap d'Or, Nova Scotia, described by Powers and Lane (1916). This flow differentiated by sinking of pyroxene and rising of plagioclase (the latter queried by Bowen). Good chemical and micrometric data are given, as the authors had available the core of a bore which passed through the entire flow. Matters have been slightly complicated by foundering of the crust, but the differentiation is very similar to that of the Palisades, except that there was no settling of olivine, and the range of silica percentage is even less. This example has recently been redescribed by Lund (1930).

Wagner (1938) has described a series of diabase sills from Thüringen, some of which show very clear gravitational concentration of pyroxene at the base, but the initial magma is of a somewhat curious composition with higher iron and alkalies but lower lime and silica than that of the Palisades. It is, nevertheless, interesting that the acid veins rich in quartz

and albite should occur near the upper contact and that the pyroxene is pigeonitic.

It will thus be seen that, while a great many of the features of the Palisades are paralleled in other diabase intrusions, none of them shows the curious stratum of olivine diabase.

GENERAL CONCLUSIONS

The following points are now regarded as being established:

(1) The main differentiation of the sill was effected by the settling of early-formed olivine crystals followed at a later stage by pyroxene.
(2) Although the sinking of the two minerals probably overlapped to some extent, crystallization of olivine ceased at an early stage, when the magma was still fluid, whereas pyroxene continued to form and to sink until brought to rest by increasing viscosity of the magma and by crystal interference.
(3) There was no appreciable sinking of iron ore or of calcic plagioclase though a certain amount of the latter was carried down attached to pyroxene.
(4) The contacts of the olivine diabase layer are invariably gradational, and this layer represents the fall of a shower of small olivine crystals probably precipitated in the labile region.
(5) Highly quartzose diabase is confined to veins and schlieren, principally in the upper portion of the sill, and Lewis' data overestimate the quantity of quartz in the sill.
(6) The later stages of crystallization were characterized by the segregation of "wet" pegmatitic veins and schlieren chiefly in the upper parts of the sill, followed by pronounced hydrothermal activity.
(7) The last magmatic residues were highly sodic and were injected into the marginal portions of the sill to form thin quartz-albite-pyroxene veins.
(8) The average chemical composition of the sill agrees well with that of the lower basaltic contact, and there is no evidence for introduction of silica by syntexis with floated arkose slabs.
(9) Late-formed hypersthene is abundant in the lowest third of the sill, while pigeonite is common in the upper two thirds; these two pyroxenes are antipathetic.
(10) Apart from the possibility of a certain upward migration of volatiles and resultant hydrothermal alteration, simple crystal fractionation accounts adequately for all the differentiation phenomena displayed by the sill.

Other suggestions rest on a less secure basis, and it is obvious that many points require more detailed investigation. If this contribution should serve as a stimulus for further work on the sill, one of its chief objects will have been accomplished.

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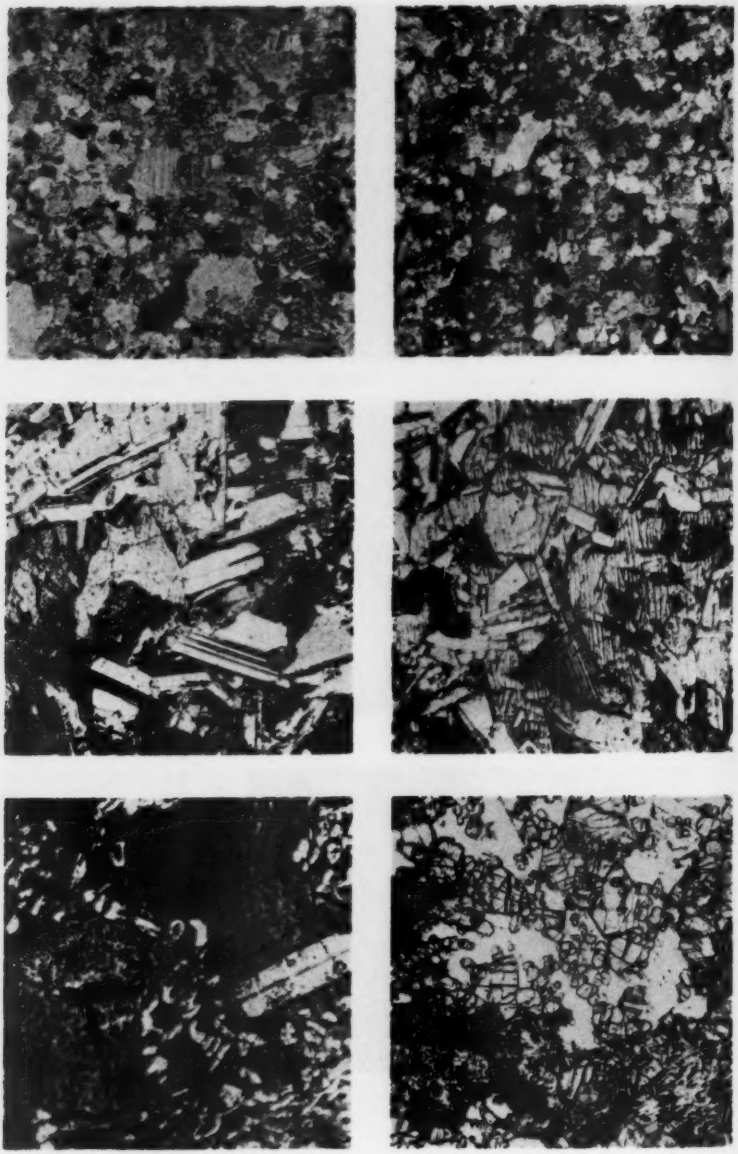
FIGURE 1. EDGEWATER CLIFF ROAD SECTION
Olivine diabase layer shows marked thinning out to left of pole.

FIGURE 2. EDGEWATER CLIFF ROAD SECTION
Joint columns pass unbroken through dark layer of olivine diabase. Hard bands in diabase coincide with joints.

FIGURE 3. ENGLEWOOD CLIFF ROAD SECTION
Light residual injection vein traversing diabase 20 feet above lower contact.

FIGURE 4. RIDGE BETWEEN FAIRVIEW AND DALLYTOWN AVENUES
Light residual injection vein traversing diabase 50 feet below upper contact.

VIEWS OF THE PALISADE SILL.



PHOTOMICROGRAPHS OF ROCKS OF THE PALISADE SILL

Explanation of Plate 2

Photomicrographs of Rocks of the Palisade Sill

Figure

1. Residual injection vein, 20 feet above lower contact, Coytesville. A granular mosaic of albite, pyroxene and quartz. Crossed nicols. x 18.

2. Arkose, 3 feet below lower contact, Coytesville. An aggregate of angular grains of quartz, soda and potash feldspar, pyroxene and iron ore. Crossed nicols. x 18.

3. Diabase, 500 feet above lower contact, electric railroad track, Fort Lee. A crystal of pyroxene is seen to the left of the center with a dark central portion of uniaxial pigeonite and a lighter margin of hypersthene-augite ($2V = 47°$). Crossed nicols. x 18.

4. Diabase, 10 feet below olivine-diabase layer, George Washington Bridge section. The bulk of the field is occupied by an ophitic crystal of hypersthene. Crossed nicols. x 18.

5. Diabase, 50 feet above olivine-diabase layer, George Washington Bridge section. Graphic intergrowth of clinopyroxene in hypersthene. Crossed nicols. x 130.

6. Olivine-diabase, middle of olivine-diabase layer, George Washington Bridge section. Packing of small olivine grains between larger constituents. Ordinary light. x 18.